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THE STRATIGRAPHY, PALAEOLOGY AND SEDIMENTOLOGY OF THE
UPPER PENDLEIAN AND LOWER ARNSBERGIAN OF THE NORTH
STAFFORDSHIRE BASIN AREA.

by
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INTRODUCTION.

Recent work by B.K. Holdsworth (Thesis 1963) has shown that within the north Staffordshire basin area a marked change in clastic deposition takes place in the Upper Pendleian and Lower Arnsbergian. This change is from 'Calcareous Siltstone' to 'Protoquartzitic Turbidite' deposition. The part of the succession including this major change in clastic deposition was suggested to the author as being suitable for study by Dr. B.K. Holdsworth (University of Keele) and it was decided to trace the strata over the whole of the north Staffordshire basin, and to compare them briefly with surrounding areas.

The object of the research was to establish in detail the palaeontological succession utilising mainly the thicker shelled goniatites, and to study the sedimentology of the coarser grained lithofacies.

During the fieldwork the discovery of seven K-bentonite beds useful for correlation provided an additional topic as also did the discovery of two Eumorphoceras horizons not previously described from the area. An interesting variation in Carboniferous cyclic sedimentation was observed within the basin area, and this is compared with some other described Carboniferous cycles.

ACKNOWLEDGEMENTS.

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ABSTRACT.

The detailed palaeontological and lithological succession was established by detailed measurement and mapping of the Hurdlow, Pyecclough and Blake Brook sections. Faunas were found and are described ranging from E₁c to E₂b₂ in age, and their extent within the north Staffordshire basin area and in Edale, Derbyshire, has been determined. The sequence of faunas described is as follows:-

E₂b₂ Eumorphoceras bisulcatum aff. leitrimense Yates
and Cravenoceratoides aff. nitidus (Phillips).

E₂b₁ { Cravenoceras cf. C. subplicatum. Bisat.
Ct. edalensis. (Bisat).
C. subplicatum. Bisat.

E₂a₃ Eumorphoceras sp. nov.II.
Leiopteria longirostris Hind.

E₂a₂ E. bisulcatum erinense Yates. E. bisulcatum
ferrimontanum Yates.

E₂a₁ E. bisulcatum grassingtonense Dunham and
Stubblefield and Eumorphoceras sp. nov.I.

E₁c C. cf. C. malhamense (Bisat).

Apart from the faunal marker horizons, seven thin K-bentonites consisting of mixed layer mica/montmorillonite with some kaolinite were found between the E₁c and E₂b₂ faunal bands (Trewin 1968). The

K-bentonites have been traced from within the north Staffordshire area to Edale, Derbyshire and are recognised in the logs of the Ashover boreholes (Ramsbottom et al. 1962).

The sedimentology of the coarser grained lithofacies is considered in detail. The protoquartzites are considered to be the deposits of turbidity currents which flowed generally northwards in the north Staffordshire basin. The deposits of the currents can be traced from a 'proximal' to a 'distal' area of deposition relative to the source area of the currents, which was probably a delta complex bordering a land mass to the south.

The calcareous siltstones contain sedimentary structures which differ from those in the protoquartzites and the beds do not appear to have been deposited by the same type of current that deposited the protoquartzites. The calcareous siltstones are never the less, considered to have been deposited by dilute suspension currents.

Cyclic sedimentation recognised in the basin area consists of cycles starting with a marine band passing upwards through calcareous siltstones, shale-mudstone, protoquartzites and shale-mudstone to the next marine band. The cycles are compared with 'Yoredale' and

'Coal Measures' cycles, and the presence of cycles in the basin is explained in terms of delta formation on a nearby shelf in response to periods of transgression and regression.

The succession deposited in Edale to the north of the Derbyshire limestone 'massif' area was found to lack the coarse grained protoquartzites but to retain most other features of the Staffordshire succession. A thin succession at Astbury is interpreted as lying on the western margin of the north Staffordshire basin in a shallow water area out of the reach of turbidity currents.

HISTORY OF RESEARCH.

The part of the succession studied has received little specific attention in the past largely because of the difficulty of elucidating the detailed succession before the advent of detailed goniatite chronology.

In 1864 Hull and Green produced a map of north Staffordshire and recognised the following succession in the Carboniferous.

Millstone Grits

Yoredale Quartzites

Shale and thin limestones

Carboniferous limestone.

It is obvious that the protoquartzitic turbidites of the present work constitute part of the Yoredale Quartzites of Hull and Green (1864) since the authors clearly indicate that they occur on Bosley Minn, Gun Hill and Lask Edge.

Hind and Howe (1901) included the strata described in the Pendleside Group, and Gibson and Hind (1899) described the succession at Astbury and give a faunal list for the marine beds in the ganister quarry (Hd. proteus horizon). Hind (1902) also lists the fauna from the ganister quarry and also records other faunas in the area, although the exact horizons cannot usually be deduced from the lists.

Pocock et al. (1906) show the anticlines of Gun Hill, Bosley and Wincle Minns, Lask Edge and Croker Hill all to consist of 'Crowstones' within the Pendleside Series. Holdsworth (1964) has discussed the extensive and ambiguous use of the term 'Crowstone' in the local geology. Pocock was clearly referring to the proto-quartzites when using this term, but Hudson and Cotton (1943) use the term Alport Crowstones to describe a calcareous siltstone type of lithofacies, and the Crowstones of Hudson (in Hudson and Cotton 1945a p.321) in the Gun Hill borehole are also predominantly of calcareous siltstone type.

Challinor (1929) recorded E. bisulcatum from his locality 521 which from his map appears to lie just above the top of the Morridge Grits which are equivalent to the protoquartzites of the present work. He mentions that the lower part of the Churnet shales in which his E. bisulcatum occurs looks very like the 'Limestones and Shales' below the Morridge Grits. The structure in this region is more complex than envisaged by Challinor and this locality — which is probably in the E2a2 faunal band is actually lower stratigraphically than parts of the area he mapped as Morridge Grits. This locality has not been located during the present work.

The Gun Hill borehole provided useful information on the succession in Staffordshire which was divided

as follows by Hudson (in Hudson and Cotton 1945a).

Churnet	Lower <u>Reticuloceras</u>
Shales	<u>Homoceras</u>
	Upper <u>Eumorphoceras</u>
1. Sandstones	{ Sandstone of Old Hay Top Position of <u>E. bisulcatum</u> Thorncliff Sandstones
Morridge Grits	2. Barren Sandy or Calc. Crowstones shales + thin quartzitic siltstones.
	Tentative position of <u>C. leion</u>
	3. Shelly Sandstones — Onecote Sandstones

Hudson also recorded a fauna with Eumorphoceras from the Blake Brook section and correlated it with the E. bisulcatum fauna in Edale at Barber Booth (Hudson and Cotton 1945) which had previously been recognised by Jackson (1927). Jackson also recognised the unconformable relation of the Edale shales (E—Lr.R1 in age, Jackson 1927 Fig.2 P.17) to the Carboniferous Limestone and also recognised a horizon with Posidonia membranacea below the E. bisulcatum horizon. Hudson noted the unconformable relationship that exists in the Upper Dove between the Namurian shales and the Carboniferous Limestone.

Fossiliferous localities within the 'E zone' were recognised by several authors, but the first full

description of the palaeontological and stratigraphical succession in any area is due to Holdsworth (thesis 1963) working in the Longnor-Hollinsclough-Morridge area.

He revised the stratigraphical nomenclature as follows:-

Churnet Formation	{	
		Thorncliff Sandstone Member

Gun Hill Siltstone Formation

Onecote Sandstone Formation

In the Upper Dove valley the coarse grained lithologies are absent and the E2 part of the succession is included in the Dove Shale Formation of Holdsworth (thesis 1963 p.175) which is roughly equivalent to the Edale Shales (Jackson 1927) to the north of the limestone massif. In the southern part of the area studied by Holdsworth coarser grained sediments appear in the succession and Holdsworth used the term Manifold Formation for the equivalent of the Churnet Formation in this area.

The faunal succession and stratigraphical formations are shown in Holdsworth (1966, Text Fig.1) and the three separate units of protoquartzitic turbidites recognised by the author are shown diagrammatically in the top of E1 and in E2a. In the south of the area studied recent work by Morris (thesis 1966) has contributed to the palaeontological knowledge of the area particularly in the E1 part of the succession, but the author does not

agree with the interpretation of the E2 faunal succession given by Morris. This will be referred to when appropriate.

Evans et al. (1968), in the Institute of Geological Sciences memoir on the Macclesfield area, have introduced further new names to the stratigraphical nomenclature by calling the lowest part of the Namurian the Lask Edge Shales. These beds — which they refer to as mudstones in the text — range from their presumed base of the Namurian up to "the base of the strata of crowstone facies" (Ibid p.20). The succeeding beds are termed the Minn Beds and are said to consist of 'alternations of mudstone and sandstone of crowstone type'. From this it seems that the base of the Minn Beds coincides with the first entry of protoquartzites into the basin area, which is the base of Unit A of the succession given in Fig.1.1a of this thesis. The top of the Minn Beds is defined by Evans et al. (1968 p.20) at the base of the Cravenoceratoides edalensis faunal band and thus the Minn Beds include Units A B and C of protoquartzites and units β and γ of calcareous siltstones (see Fig.1.1a). The base of the Minn Beds is defined at a lithofacies change and the top at a faunal band. Above the Ct. edalensis band the strata are termed Churnet Shales by Evans et al. (1968).

Within the basin area, which includes both the Minns and Lask Edge, the limits of the Lask Edge Shales and

Minn Beds can be correlated with the present work. It should be noted that the Minn Beds contain strata of both the 'calcareous siltstone' and 'protoquartzite' lithofacies of Holdsworth (1963, 1963a) and that the 'Lask Edge Shales' consist of calcareous siltstone lithofacies. In the Gun Hill borehole (Hudson in Hudson and Cotton 1945a) the equivalent beds of the Lask Edge Shales were termed 'crowstones' and are of calcareous siltstone lithofacies (Holdsworth thesis 1963). Thus to state that the Minn Beds begin at the base of strata of 'crowstone' lithofacies is confusing since this term has in the past been applied to both 'protoquartzite' and 'calcareous siltstone' lithofacies (Holdsworth 1964). Despite this confusion over the use of the term crowstone it seems clear that in the basin area the division between the Lask Edge Shales and Minn Beds coincides with the junction between Unit α and Unit A of the succession described in this thesis.

Turning now to the Astbury area, which is not within the basin area, a serious discrepancy arises between the conclusions of Evans et al. (1968) and the author. Evans et al. (1968) state that the Lask Edge Shales 'are best exposed in Limekiln Brook, Astbury, but to avoid confusion with the underlying Astbury Limestone-Shales they are here named the Lask Edge Shales'. Thus they appear to take the type section

as being in Limekiln Brook but derive the name for the formation from a locality to the east in the basin area. In the Limekiln Brook section useful goniatites are absent and hence there is great uncertainty about the age of the exposed strata. The writer considers that the divisions of Minn Beds and Lask Edge Shales given by Evans et al. (1968 Fig.6) for the Limekiln Brook section do not correspond with those given for the same formations on Lask Edge or in the Gun Hill borehole. The Lask Edge Shales of Limekiln Brook section contain near their base thin developments of calcareous siltstone lithofacies alternating with mudstones with sideritic sandstones. These alternations are thought by the writer to represent the alternations seen in the Lask Edge Shales and the Minn Beds of Lask Edge and the Minns (i.e. Units α , β , γ and A,B,C of the succession given in this thesis). The fauna with Posidoniella variabilis at Loc.23 of Evans et al. (1968) is considered by the author to be of E2b age, and thus the whole of E1 and E2a are included by the author in the lower portion of the Lask Edge Shales at Limekiln Brook whilst on Lask Edge and the Minns the Lask Edge Shales are of E1 age and the Minn Beds range from E1 to the base of E2b.

Generally speaking the Minn Beds of the Minns correlate with the Yoredale Quartzites of Hull and Green (1864); the Morridge Grits of Challinor (1929); the

Thorncliff Sandstones with the Old Hay Top sandstone of Hudson (in Hudson and Cotton 1945a), and the Thorncliff Sandstone Member of Holdsworth (thesis 1963).

For the purpose of this thesis the author has avoided the introduction of new names and as far as possible use of the existing ill-defined names. The succession has been divided into units as shown in Fig.1.1a. The majority of this thesis is concerned with the strata from the bases of Unit A which underlies the C. cf. C. malhamense E1c faunal band up to the Cravenoceratoidea edalensis E2b1 band. The units described have been traced over the whole of the north Staffordshire basin area and into Edale, Derbyshire. The only section where they cannot be recognised with certainty is the Limekiln Brook section at Astbury. It is thus unfortunate that this locality was chosen as a 'type' section by Evans et al. (1968). The author's interpretation of the Astbury section is discussed further in part 3.1 of this thesis.

The lack of sedimentological conclusions relating to the area can be ascribed to the generally imperfectly known succession. Challinor (1928) recognised the broad transition from marine limestones at the base to coal swamp conditions at the top of the Carboniferous succession in the Morridge area. In the early memoirs of the area (Pocock 1906, Gibson and Wedd 1902) environmental interpretations are made particularly concerning the

presence of coal seams at Astbury and plant remains in the sandstones. Studies not directly concerned with the present work have been made in north Derbyshire by Allen (1960) and Walker (1966) which illustrate turbidite facies associated with the southerly advancing deltas of R1.

Holdsworth (thesis 1963: 1963a) was the first to recognise the range of lithologies to be found in the Namurian of north Staffordshire. He recognised the turbidite nature and southerly derivation of the proto-quartzitic turbidites and showed that they could be distinguished from the calcareous siltstone lithofacies. He also recognised that the calcareous siltstones do not occur higher than E2a in the north Staffordshire succession, and that the Viséan massif and basin areas of sedimentation continued to be important in the Lower Namurian (Holdsworth 1963a).

The present work extends Holdsworth's study of the 'Protoquartzite' and 'Calcareous Siltstone' lithofacies to cover the whole of the exposed part of the basin area for the Upper Pendleian and Lower Arnsbergian.

The nature of the marine band sediments and faunas as preserved in bullion limestones in the Namurian succession has been described by Holdsworth (1964a, 1966) using material ranging upwards in age from E2b.

PART 1

THE GENERAL STRATIGRAPHY AND MARKER HORIZONS.

SECTION 1.1

INTRODUCTION AND ESTABLISHMENT OF SUCCESSION.

The part of the succession studied constitutes part of the Upper Pendleian (E1) and Lower Arnsbergian (E2) stages of the Namurian. The general succession in north Staffordshire is shown in Fig. 1.1a and consists of alternations of several different lithofacies which will be discussed in detail when the sedimentology is considered. The author's interpretation of the succession in Edale, Derbyshire is shown in Fig. 1.1b. The two major coarse grained lithofacies are referred to as the 'protoquartzites' and the 'calcareous siltstones'. There are three units of protoquartzites which are referred to as A, B and C and three units of calcareous siltstones which are termed α , β and γ for convenience. Also present is a unit consisting of shale mudstone with siderite nodules. Shale and shale-mudstone is found between individual beds of both of the coarse grained lithofacies, and is also the major constituent of the marine horizons.

There are 4 marine horizons with goniatites that are of major importance in correlation. These are

Fig 1.1a

GENERAL STRATIGRAPHY OF THE U. PENDLEIAN AND L. ARNSBERGIAN OF NORTH STAFFORDSHIRE.

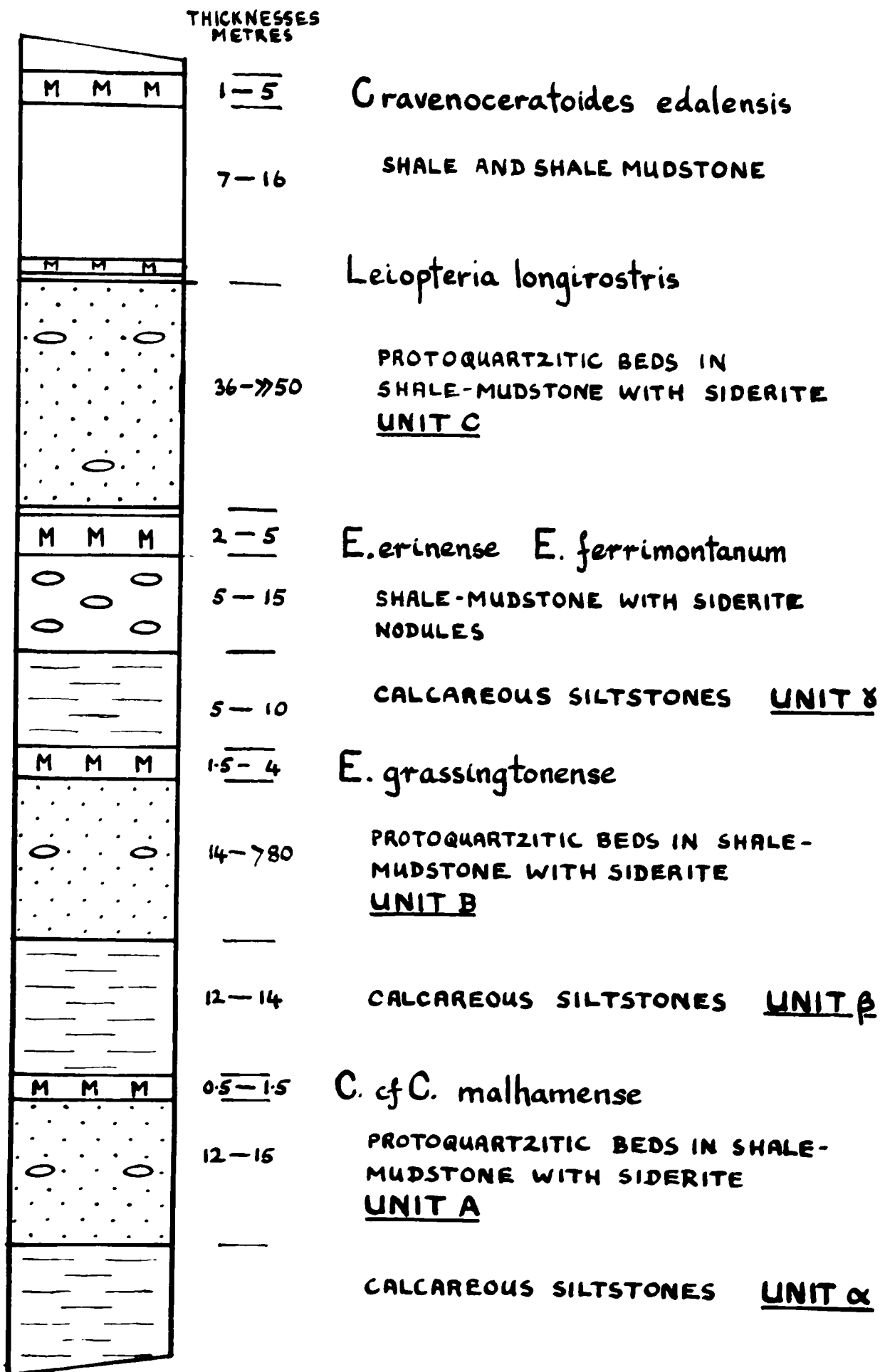
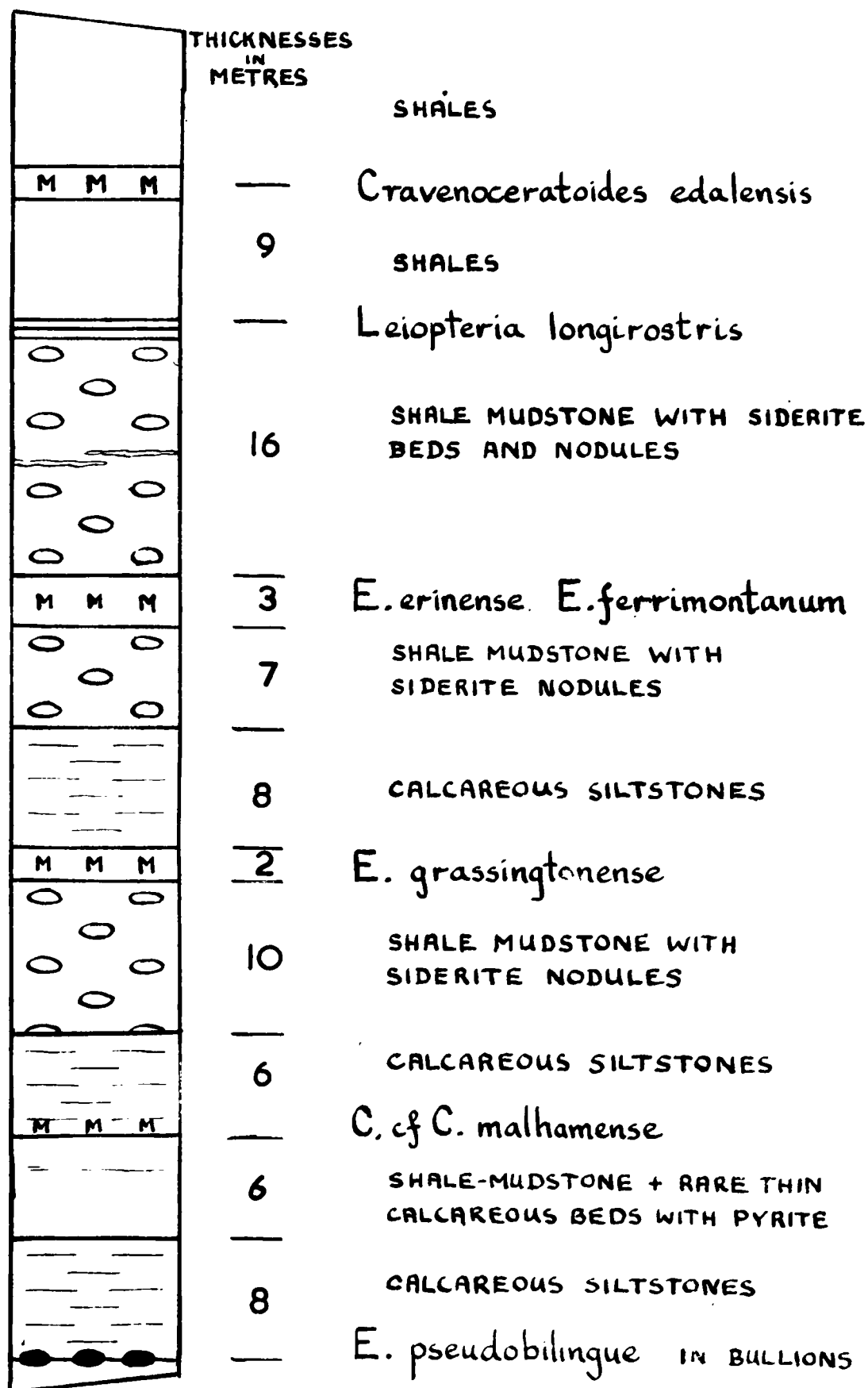


Fig 1.1b

GENERAL FAUNAL AND LITHOLOGICAL SUCCESSION IN EDALE, DERBYSHIRE.



1) C. cf. C. malhamense (Bisat) E1c; 2) E. bisulcatum grassingtonense Dunham and Stubblefield E2a1;
 3) E. bisulcatum ferrimontanum Yates and E. bisulcatum erinense Yates, E2a2; and 4) Ct. edalensis (Bisat) E2b1. In addition to these there are several other goniatite horizons in the region of the Ct. edalensis horizon that are useful within the present area, and also a horizon which appears to be quite widespread with the distinctive lamellibranch Leiopteria longirostris Hind. All these horizons are described in the following section. Since the area is so large no attempt has been made to map it systematically, but since the divisions of calcareous siltstones and protoquartzites have not been precisely recognised before and do not as yet appear on any published map of the area three large scale geological sketch maps are included of three of the localities where all the sedimentary units and faunal bands can be seen. These maps are of the Hurdlow, Blake Brook and Pyeclough sections (Figs. 1.1c, 1.1d, 1.1e).

The terminology in current use in the literature with regard to Eumorphoceras is very confused. Various 'forms' of Eumorphoceras have been described as subspecies, varieties or mutations of E. bisulcatum Girty. The author therefore has attempted to evaluate the 'forms' relevant to the present work with reference

to E. bisulcatum Girty, and has decided to elevate the subspecies erinense, ferrimontanum, leitrinense and grassingtonense to specific rank. The discussion relevant to this procedure follows.

Explanation Figs.1.1c,d,e.

Sketch maps of the Pyeclogh, Blake Brook and Hurdlow stream sections to illustrate the succession. Calcareous Siltstones, Units α , β and γ shown in blue, and protoquartzite Units A, B and C shown in red. Major marine bands are also indicated.

- E2b1 - Ct. edalensis
- E2a2 - E. erinense and E. ferrimontanum
- E2a1 - E. grassingtonense
- E1c - C. cf. C. malhamense

Fig 1.1c

PYE CLOUGH

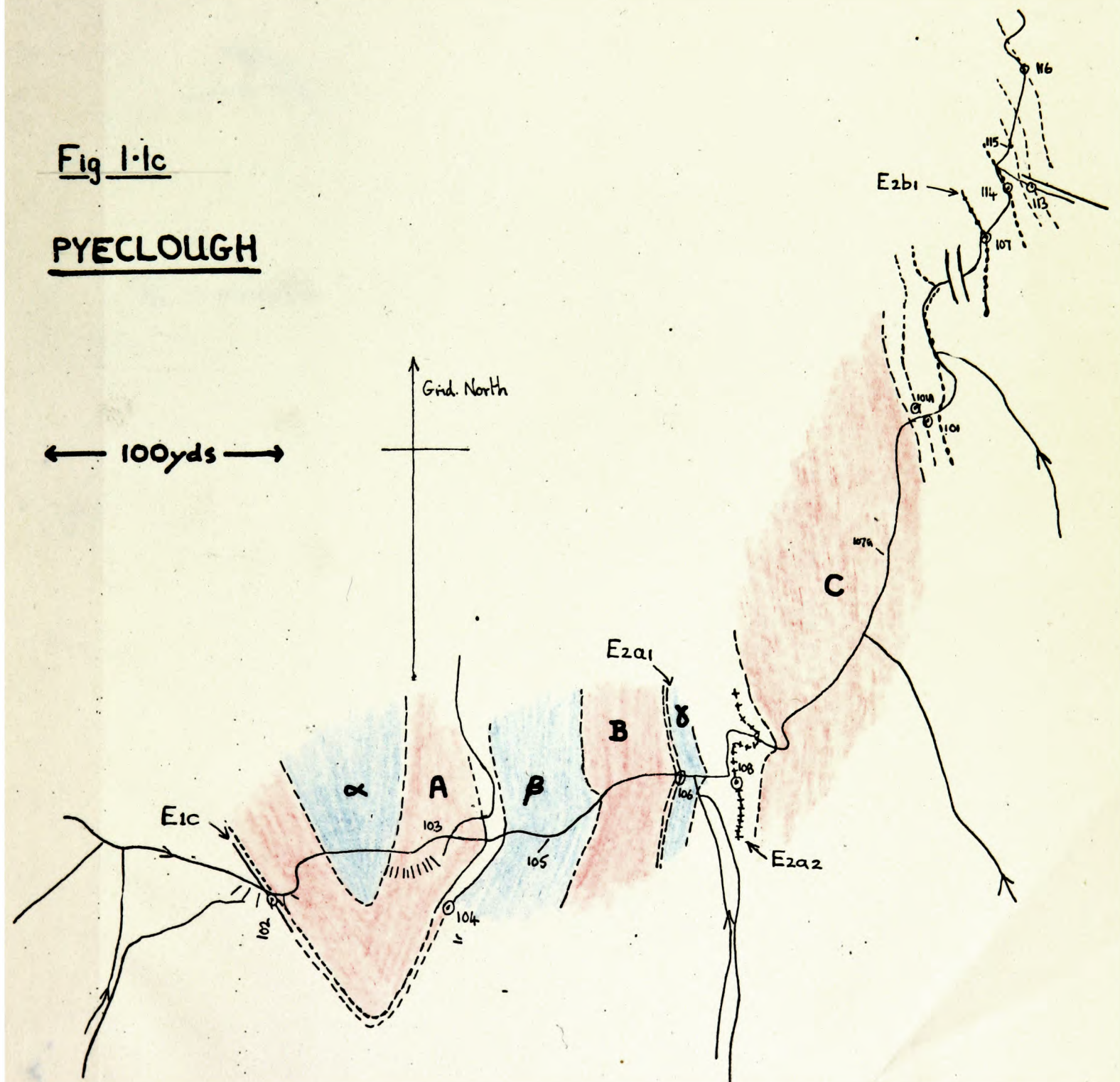


Fig 1.1d

BLAKE BROOK

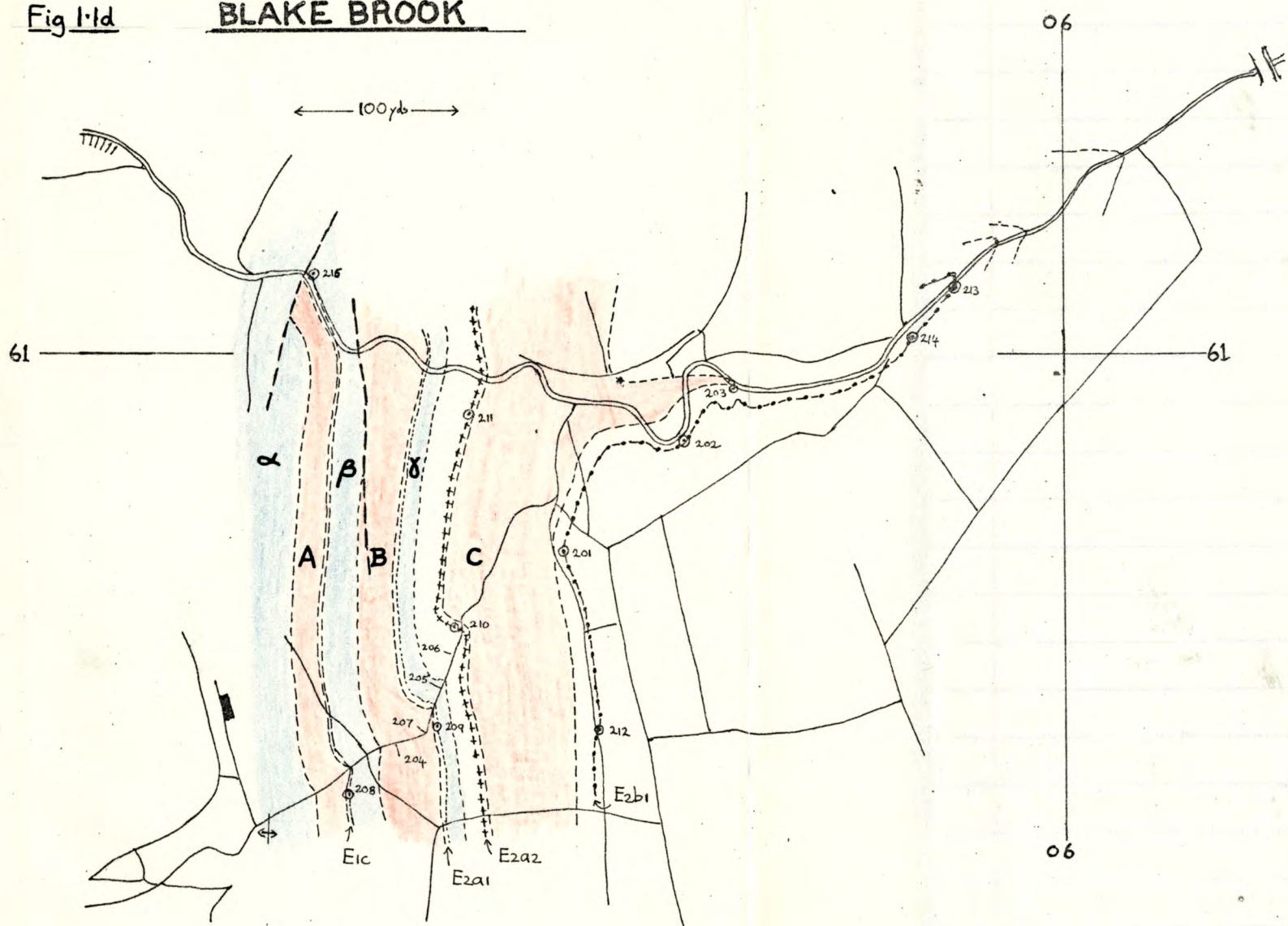
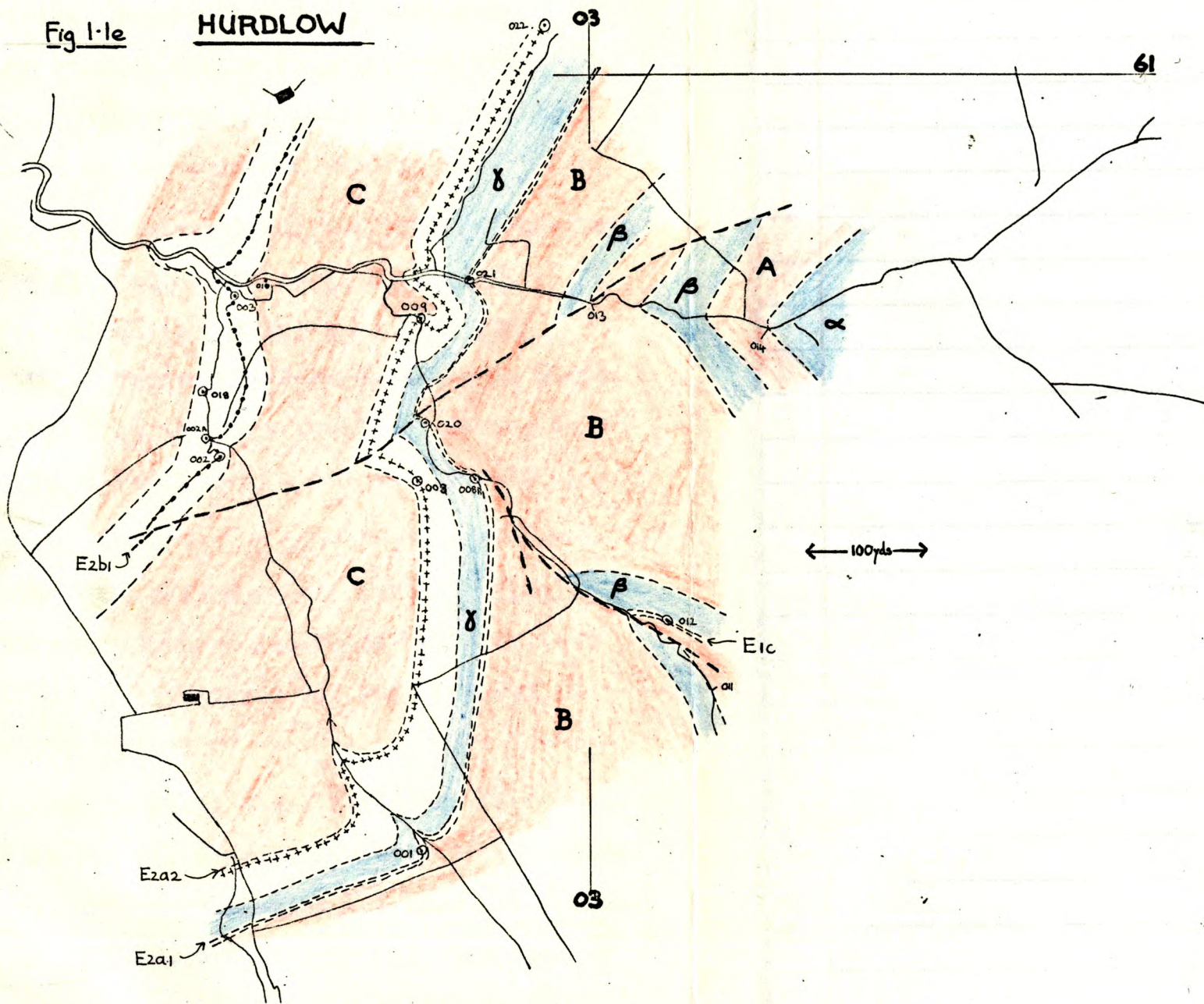


Fig 1-1e

HURDLOW



SECTION 1.2

DISCUSSION OF EUMORPHOCERAS BISULCATUM Girty.

In his original description Girty obviously interpreted this species rather broadly, and many English species of Eumorphoceras would fall within this species based on the original description.

Part of the type material has now been redescribed, and one specimen figured by him as E. bisulcatum (1909 Pl.XI Fig.16-16b) is now known as E. girtyi Elias (1956). This specimen differs markedly in shape and number of ribs from the type specimen, having only 17 ribs on the final whorl seen at 10 mm. overall diameter.

The separation of this specimen from E. bisulcatum leaves only three specimens illustrated by Girty that are referred to as E. bisulcatum and one of these is very juvenile.

The type specimen Pl.XI Figs.15-15c has 18-19 ribs / $\frac{1}{2}$ whorl at 7.6 mm. overall diameter (6 mm. Spiral Groove diameter), and about 17 ribs / $\frac{1}{2}$ whorl at 9.0 mm. overall diameter (7.6 mm. S.G. diameter). The size of the umbilicus is 2.5 mm. and c3.0 mm. at the above diameters respectively. Fig.17 Pl.XI shows a specimen with c17 ribs / $\frac{1}{2}$ whorl at 6.3 mm. overall diameter

(5.6 mm. S.G. diameter) and an umbilicus 2.3 mm. wide. The later part of the final whorl in the type specimen is poorly preserved and the shell is partly missing. Neither the full shape of the ribs nor their number can be accurately measured, but ribs are still present at the largest diameter seen.

The ribs are sickle shaped and extend over the flanks as far as the spiral groove. For the first $\frac{2}{3}$ of their passage over the flanks the ribs are straight or form a slight forward bow. The ribs are not radial but are directed slightly adorally. In the last $\frac{1}{3}$ the ribs bend sharply forward to meet the spiral groove, and become reduced in strength until they merge with the spiral groove. The ribs do not always appear to be regularly spaced in the type material. Fig.17 (Girty 1909) shows irregular spacing and strength of ribs at only 6 mm. overall diameter, and the type specimen shows the feature of one rib rising a little way from the umbilical edge and very close to the preceding rib. This last feature is best seen in the specimen illustrated by Gordon (1964, Pl.25 Fig.18).

The diameter at which the ribs start to die out is crucial in identifying the English Eumorphoceras specimens, and it is unfortunate that this diameter is not accurately known in the case of E. bisulcatum Girty.

McCaleb, Quinn and Furnish (1964) state that the prominent lateral ribs become indistinguishable at an early intermediate stage - 25-30 mm. diameter. Their Fig.5, PL.2 shows an internal mould of a septate specimen considered to be E. bisulcatum and there appear to be small nodes present near the umbilicus at 30 mm. (23 mm. spiral groove) diameter, and at 24 mm. (19 mm. spiral groove) diameter ribs appear to be present on the flanks but are only faintly seen on the internal mould. Thus it seems probable that the ribs reduce to umbilical nodes between these diameters. At larger diameters the shell is apparently smooth apart from growth lines.

Summary description of essential features.

E. bisulcatum Girty must now be restricted to a Eumorphoceras with strong ribs still present at 11 mm. overall diameter (9 mm. S.G. diameter). If the material of McCaleb, Quinn and Furnish is conspecific with the type, then the ribs die out to umbilical nodes between 19 mm. and 23 mm. S.G. diameters. The ribs number 34-38 per whorl between 6 mm. and 10 mm. overall diameter and are sickle shaped. At 11 mm. overall diameter (c9 mm. S.G.) the umbilicus is about 3.5 mm. wide. These features are useful in identifying material whether it be solid or crushed. Comparisons

of suture or whorl shape cannot be made with the present crushed material.

The Introduction of *E. bisulcatum* to the British Literature.

Bisat (1924a, b) first introduced *E. bisulcatum* to the British Literature and used it in the broad sense in which it was originally used by Girty. Bisat included the forms found in E2a and considered that the *Eumorphoceras* associated with *Nuculoceras nuculum* to be "the same species (or a very close ally)".

In view of the variation in Girty's original material this was permissible, but with the separation by Elias of Girty's original material the definition of *E. bisulcatum* becomes more restricted as is outlined above. Thus the features of the British and Continental specimens that have been included within *E. bisulcatum* must be reconsidered with reference to the type material.

Notes on the differences between *E. bisulcatum* Girty and European subspecies of *E. bisulcatum*.

From Fig. 1.2a it can be seen that the only subspecies of *E. bisulcatum* that in any way resemble *E. bisulcatum* Girty are the subspecies *erinense* and *ferrimontanum* which have ribs of comparable shape and strength. These subspecies both differ from *E. bisulcatum* Girty in having significantly fewer ribs at 6 mm. S.G. diameter and also in that from this diameter the rib

Fig 1.2a

	Diameter Overall	Spiral Groove	No. of Ribs $\frac{1}{2}$ whorl	Diameter at which ribs die out	Umbilical - Diameter	Spiral Groove Diameter	S.G.Diam. Umb.Diam.	
<u>E. bisulcatum</u>		5.5	14		3.0	6.3	2.1	Ribs sickle shaped very similar to <u>E. bisulcatum</u> Girty
<u>erinense</u>		6.0	15		3.5	8.6	2.5	
Yates 1962		9.0	16	C.15 mm.S.G.Diameter	4.2	10.0	2.4	
		11.0	17		4.4	10.9	2.5	
		14.0	16					
<u>E. bisulcatum</u>		7.8	16		3.8	9.6	2.5	Similar rib shape to <u>erinense</u> but weaker over flanks and having marked umbilical nodes when ribs fade.
<u>ferrimontanum</u>		9.6	16	C.10 mm.S.G.Diameter	4.2	9.7	2.3	
Yates 1962		10.8	17		4.5	10.8	2.4	
		12.0	15		5.3	12.0	2.3	
<u>E. bisulcatum</u> mut.		7.5	13-16		3.7	7.5	2.0	Rib bifurcation occurs before ribs die out to umbilical nodes. Ribs thicken over mid flank region.
<u>grassingtonense</u> .		9.5		C.12 mm.S.G.Diameter	3.9	8.0	2.0	
Dunham & Stubblefield		12.0	10		4.4	9.5	2.2	
1944					5.0	11.0	2.2	
<u>E. bisulcatum</u>		6.3	9	Ribs do not reach	1.5	4.0	2.6	Strong constrictions - ribs short and stubby not reaching groove.
<u>leitrimense</u>		6.4	10	$\frac{1}{2}$ way to S.G. from	2.1	6.2	3.0	
Yates 1962		6.5	11	6.4 mm.S.G.diam. Smooth at 10 mm.Diam.	2.8	7.6	2.7	
<u>E. bisulcatum</u>	10.7	19.3	15		4.7	9.3	2.0	Short ribs - do not reach spiral groove - constrictions present.
<u>varicata</u> Schmidt 1934	12.7	11.0	13	Not known	5.3	11.0	2.1	
<u>E. bisulcatum</u>	14	10	c 9	Not known	3	10	3.3	Ribs grouped in pairs - spiral ornament on venter.
mut Schmidt 1934								
<u>E. bisulcatum</u> s.l.	1.5		9		3.3	8.2	2.5	Occasional rib with umbilical end supressed - spiral ornament on venter - similar to <u>mut</u>
Moore. 1946	3.0		8-9					
	10.0	8.2	12	ribs present on flanks				
	12.0		13	at 27 mm.S.G.diam.				
	7.0	6.0	11		2.0	6.0	3.0	
	31.0	27.0	c 13		6.5	27.0	3.4	
<u>E. bisulcatum</u>	11	9			c 3.5	9	2.6	Sickle shaped ribs reaching spiral groove.
Girty 1909	6.3	5.6	17		2.3	5.6	2.4	
	7.6	6.0	18-19	19-23 mm. S.G.	2.5	6.0	2.4	
	9.0	7.6	17	diameter	c 3.0	7.6	2.5	

counts of erinense and ferrimontanum increase to $17\frac{1}{2}$ whorl at 10-11 mm. diameter whereas in E. bisulcatum Girty the count decreases to $17\frac{1}{2}$ whorl from 18- $19\frac{1}{2}$ whorl.

The diameter at which the ribs die out is also different. The apparent difference in umbilical size mentioned by Yates (1962) is due to imperfect nature of the type specimen which does not allow the umbilical diameter that corresponds with the maximum diameter of the shell to be directly measured.

Thus in the author's opinion significant differences occur between all these different described subspecies of Eumorphoceras except possibly between E. bisulcatum s.l. of Moore (1946) and E. bisulcatum mut. β Schmidt which need further investigation. They are both very different from E. bisulcatum Girty.

The author considers that with the separation of part of Girty's type material of E. bisulcatum as E. girtyi Elias, the definition of the species E. bisulcatum becomes restricted. Therefore the described subspecies of E. bisulcatum should be elevated to specific rank in view of their clear differences from E. bisulcatum as outlined above. Thus the description by Moore (1946) of E. bisulcatum s.l. is not considered as an addition to the original definition of E. bisulcatum Girty but as a description of a Eumorphoceras differing

markedly from the type material of E. bisulcatum Girty and warranting an individual specific name.

It is proposed that the subspecies erected by Yates (1962) should be elevated to specific rank and become E. ferrimontanum, E. erinense and E. leitrinense and that E. bisulcatum mut. grassingtonensis Dunham and Stubblefield (1944) should become E. grassingtonense.

The author has not examined enough material belonging to the other described groups to be certain of their detailed characteristics, they are however certainly specifically distinct from E. bisulcatum.

SECTION 1.3

THE FAUNAL MARKER HORIZONS.CRAVENOCERAS cf. C. MALHAMENSE HORIZON.

This horizon is generally about 50cm. thick and occurs above the A unit of protoquartzites in the base of the β unit of calcareous siltstones.

FAUNAL LIST.

<u>Cravenoceras</u> cf. <u>C. malhamense</u> (Bisat)	Fig.1.3a1
<u>Cravenoceras</u> sp.	Fig.1.3a2
<u>Posidonia membranacea</u> (M'Coy)	Fig.1.3b1
<u>Posidonia membranacea horizontalis</u> (Yates)	Fig.1.3b2
<u>Posidonia corrugata elongata</u> Yates	Fig.1.3k2
<u>Actinopteria persulcata</u> (M'Coy)	Fig.1.3a3

Descriptions of selected species.

Order	AMMONOIDEA	Zittel 1884
Suborder	GONIATITINA	Hyatt 1884
Super family	GONIATITACEA	de Haan 1825
Family	GONIATITIDAE	de Haan
Subfamily	HOMOCERATINAE	Spath
Genus	CRAVENOCERAS	Bisat 1928

Cravenoceras malhamense (Bisat)

Homoceras malhamense Bisat. 1924 p.106-107 Pl.I, Figs. 1,2

Pl.IX Figs. 18,19

Cravenoceras malhamense (Bisat) Bisat 1932

Cravenoceras cf. C. malhamense (Bisat) Fig.1.3a

Specimens are generally very poorly preserved as crushed ventral impressions in shale, umbilical impressions are rare. The striae are very fine, straight and non-bifurcating, and arise within the umbilicus and are usually radial. These specimens are probably Cravenoceras and are tentatively assigned C. cf. C. malhamense, but without solid specimens showing suture and whorl shape it is virtually impossible to distinguish the species of Cravenoceras (see Bisat 1930. p.28).

Discussion.

The specimens illustrated by Johnson, Hodge and Fairbairn (1962) as C. cf. C. malhamense have a marked hyponomic sinus unlike the present material and unlike the original material, but Ramsbottom (in Johnson, Hodge and Fairbairn 1962) considers that they are nearer to C. malhamense than any other Cravenoceras. The present material appears to conform to the type material better than this. Their material comes from a horizon about 70 feet above C. leion and appears to be lower in the succession than the present material which is close below the E2a bands.

Cravenoceras sp. Fig.1.3a2

One specimen only was found with very closely spaced striae which are definitely curved on the flank.

This feature distinguishes the specimen from C. malhamense. The backward deflection of the striae over the flank may indicate affinities with the later C. cowlingsense Bisat (1930), but at this size the striae in C. cowlingsense are generally fewer in number. The specimen can only be referred to as Cravenoceras sp.

Class	LAMELLIBRANCHIATA	Blainville 1816
Family	PTERINOPECTINIDAE	Newell 1937
Genus	POSIDONIA	Bronn 1828
	<u>Posidonia membranacea</u>	(M'Coy)

Posidonia membranacea (M'Coy) (oblique form)

Posidomya membranacea M'Coy (1844)

Posidomya membranacea M'Coy Hind 1901 Pl.15 Figs.18-23

Posidomya membranacea (M'Coy) Smyth 1949 p.317 Pl.XVII Fig.6

Posidonia membranacea (M'Coy) (horizontal form)

Caneyella membranacea (M'Coy) Ramsbottom 1959 p.405-406

Pl.71 Fig.14

Caneyella membranacea horizontalis Yates 1962 Pl.61 Fig.5

Posidonia membranacea (M'Coy) Demanet 1938 Pl.10 Figs.5,10,11

A consideration of Posidonia membranacea is confused by the fact that there are two distinct forms of the species and possibly some forms intermediate between the two. One form has a long hinge line with the umbo

situated near the anterior end and is referred to here as the horizontal form, and the other form has a short hinge line with the umbo roughly central in position and is referred to as the oblique form.

The original description of Posidomya membranacea by M'Coy (1844) is clearly of the oblique form; as also are all the figures in Hind (1901 Pl.15 Figs.18-23). Smyth (1949) also illustrates the oblique form under the name Posidonia membranacea. It seems therefore to be clear that the specific name membranacea was originally applied to the oblique form, and it should retain this specific name.

Examples of the horizontal form are figured by Demanet (1938 Pl.10 Figs.5,10,11), and also by Ramsbottom (1959) as Caneyella membranacea. Yates (1962 p.389) has described the horizontal form as Ca. membranacea horizontalis thus making it of subspecific rank.

Ramsbottom (1959) in his paper on the distinctions between Caneyella, Posidonia and Posidoniella states that the genus Caneyella "includes both costate and a few non-costate species with a relatively long hinge line, the umbo being usually towards the anterior end." On this basis the horizontal form belongs to the genus Caneyella, but the oblique form falls within his definition of Posidonia with regards to its hinge line. He considers Posidonia to include "non-costate species

in which the umbo is more or less centrally placed on the relatively short hinge line, or at least not less than one-third of the length of the hinge line from the anterior end."

On the question of the costation of the shells, the oblique form can be non-costate so conforming with the above definition of Posidonia or it may be faintly costate as shown by Hind (1901 Pl.5 Figs.19,21,22,23). Fig.18 of Hind (1901) of the type specimen, shows very faint costae and Fig.20 appears to be non-costate. M'Coy's description calls them "fine somewhat obscure radiating lines which pass from the umbones to the lower border." He also describes the hinge line as "short, straight" with "umbones subcentral inclining to the anterior end." All the specimens of the horizontal form appear to be costate. In this area the horizontal forms are also costate, and the oblique forms may be faintly costate or non-costate.

In order to determine the relative obliquity of the shells a graph was constructed on which the angle between the hinge line and the line from the umbo to the furthest point on the posteroventral border - here termed the 'Angle of Obliquity' - was plotted against the length of the hinge divided by the distance of the umbo from the anterior end of the hinge. Measurements made on various illustrations are included on (Fig.1.3c) along

with a selection of specimens from the C. cf. C. malhamense band. The theoretical dividing line between the genera Caneyella and Posidonia on Ramsbottom's definitions of the genera is shown as a dashed horizontal line. It is clear that there is considerable variation in both these parameters and it is possible that there is even a complete transition between the two.

Ramsbottom in transferring P. membranacea to Caneyella believed the distinctions of ornament and position of the umbo on the hinge line to "have genetic significance and to separate shells which are not congeneric." The present consideration of C. membranacea does not support this statement and it seems logical to replace C. membranacea within the genus Posidonia.

C. ricardsoni - the type species of Caneyella is strongly and regularly costate with a hinge line occupying nearly the whole anterior-posterior length of the shell. It is, in general appearance very different from P. membranacea which in its general features is closer to P. becheri the type species of Posidonia.

In the C. cf. C. malhamense band the commonest species is the oblique form of Caneyella membranacea (Fig.1.3b) which is often so abundant as to completely cover some bedding planes. The vast majority of the specimens are of the oblique form but occasional specimens of the horizontal form are also met with. This is in

marked contrast with lower horizons in the area where this species is present, since in these horizons the horizontal form (Fig.1.3b2) is abundant and the oblique form absent or very rare.

The specimens found in the C. cf. C. malhamense band have a maximum dimension of 42mm. measured from the umbo to the far point on the postero-ventral margin, and a maximum width perpendicular to their elongation of 16mm. There is considerable variation in the ratios of these parameters, (termed x and b respectively) and this is shown in Fig.1.3d, measurements taken from Hind's (1901) illustrations are also shown.

The abundance of large P. membranacea in this band is the most distinctive feature noticed in the field and since the higher bands do not contain the same form, confusion is not likely to arise.

Note. All measurements made on the shells should be treated with caution as the crushing may have caused some distortion of the shells.

COMPARISON WITH OTHER AREAS.

Slieve Anierin, Ireland.

Yates (1962) recorded a similar fauna from the C. malhamense band in Ireland, and apart from the Cravenoceras also found the same lamellibranch fauna with the exception of P. corrugata elongata which she

does not record. Kazakhoceras scaliger (Schmidt) and Chaenocardiola footii (Baily) are found in Ireland at this level but have not been seen in the present area. The horizon also occurs in the same position relative to the E. grassingtonense horizon as in the present area, but the E. pseudobilingue horizons that occur below it in Ireland have not been found in the central part of the north Staffordshire basin.

Edale, Derbyshire.

In Edale, Hudson and Cotton (1945) have recognised this horizon at several localities and the fossils they record include the following:-

C. malhamense Bisat

P. aff. corrugata (Etheridge)

P. membranacea M'Coy

P. aff. membranacea (elongate form)

The localities mentioned by Hudson and Cotton have been visited and P. membranacea was found to be abundant and reached a larger size than in north Staffordshire, up to 50mm. x dimension. No true specimens of P. membranacea horizontalis were seen by the author. The abundant form is P. membranacea and this appears to be the form referred to as 'large and elongate' by Hudson and Cotton (1945).

Alport, Derbyshire.

The fauna recorded from the Alport borehole by

Hudson and Cotton (1943) includes the following at this horizon:-

C. malhamense

Neodimorphoceras scaliger

Actinopteria sp.

Pd. membranacea (elongate form)

Chaenocardiola footii

This fauna agrees well with the fauna from other areas, and there can be little doubt that the horizons all correlate with each other. The elongate form of P. membranacea referred to by Hudson and Cotton (1943 p.167) appears to be the oblique form, or true P. membranacea, as was found to be the case at the exposures in Edale.

Lancaster Fells.

Moseley (1954) also records a similar fauna from this level in the succession, but does not describe or illustrate any of the specimens and so no meaningful comparisons can be made.

LOCALITIES IN NORTH STAFFORDSHIRE.

Below is a list of localities in the present area where the C. malhamense band is exposed. None of these localities are recorded in previous literature on the area, apart from one at Oakenclough which was found by Holdsworth (Thesis 1963) who considered the horizon to

be 'probably of E2a age', on the basis of the presence of P. corrugata elongata identified by Dr. W.H.C.

Ramsbottom.

<u>Locality No.</u>	<u>General Area.</u>
011	} Hurdlow
012	
016	
102	} Pyecclough
104	
152	} Oakencclough
165	
	- - - - - Loc 342 Holdsworth Thesis p.103
208	} Blake
215	
507	} Elkstones
521	
583	} Winkhill - Martinslow
586	

Explanation Figs.1.3a,b.

Fig.1.3a

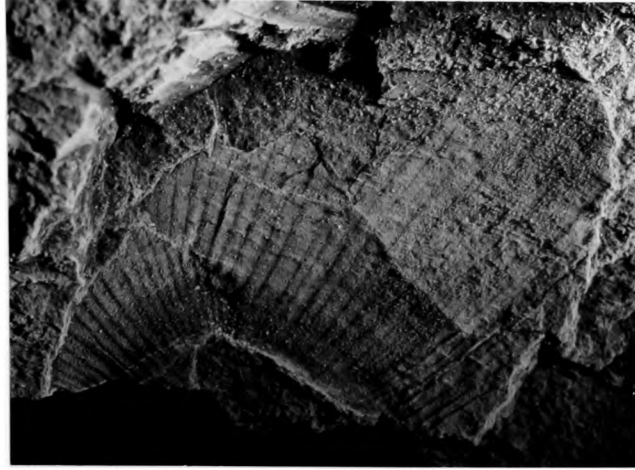
1. Cravenoceras cf. C. malhamense (Bisat) x5 Spec.104A
Loc.104 Pyecclough
2. Cravenoceras sp. x5 Spec.012A Loc.012 Hurdlow
3. Actinopteria persulcata (M'Coy) x5 Spec.012B Loc.012 Hurdlow

Fig.1.3b

1. Posidonia membranacea (M'Coy) x5 Spec.01204 Loc.012 Hurdlow
2. Posidonia membranacea horizontalis (Yates) x5 Spec.0G1
P2a horizon near Onecote Grange, N. Staffordshire.

All specimens illustrated in Figs.1.3a-b are from the E1c
C. cf. C. malhamense horizon with the exception of Fig.1.3b2.

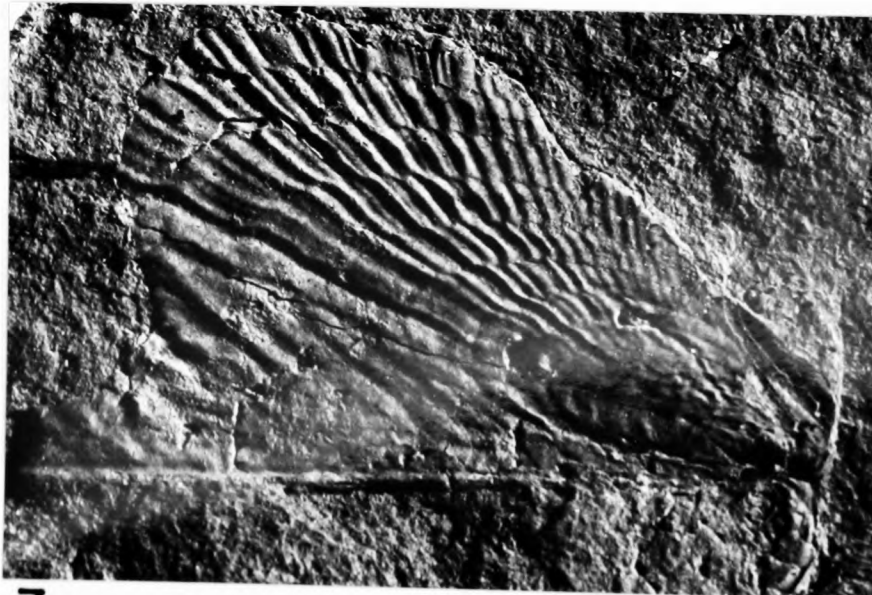
Fig 1.3a



1



2



3

Fig 1.3b



1

2



PLOT of ANGLE of OBLIQUITY and RATIO $\frac{l}{x}$ for POSIDONIA MEMBRANACEA and . P. MEMBRANACEA HORIZONTALIS.

Figl3c

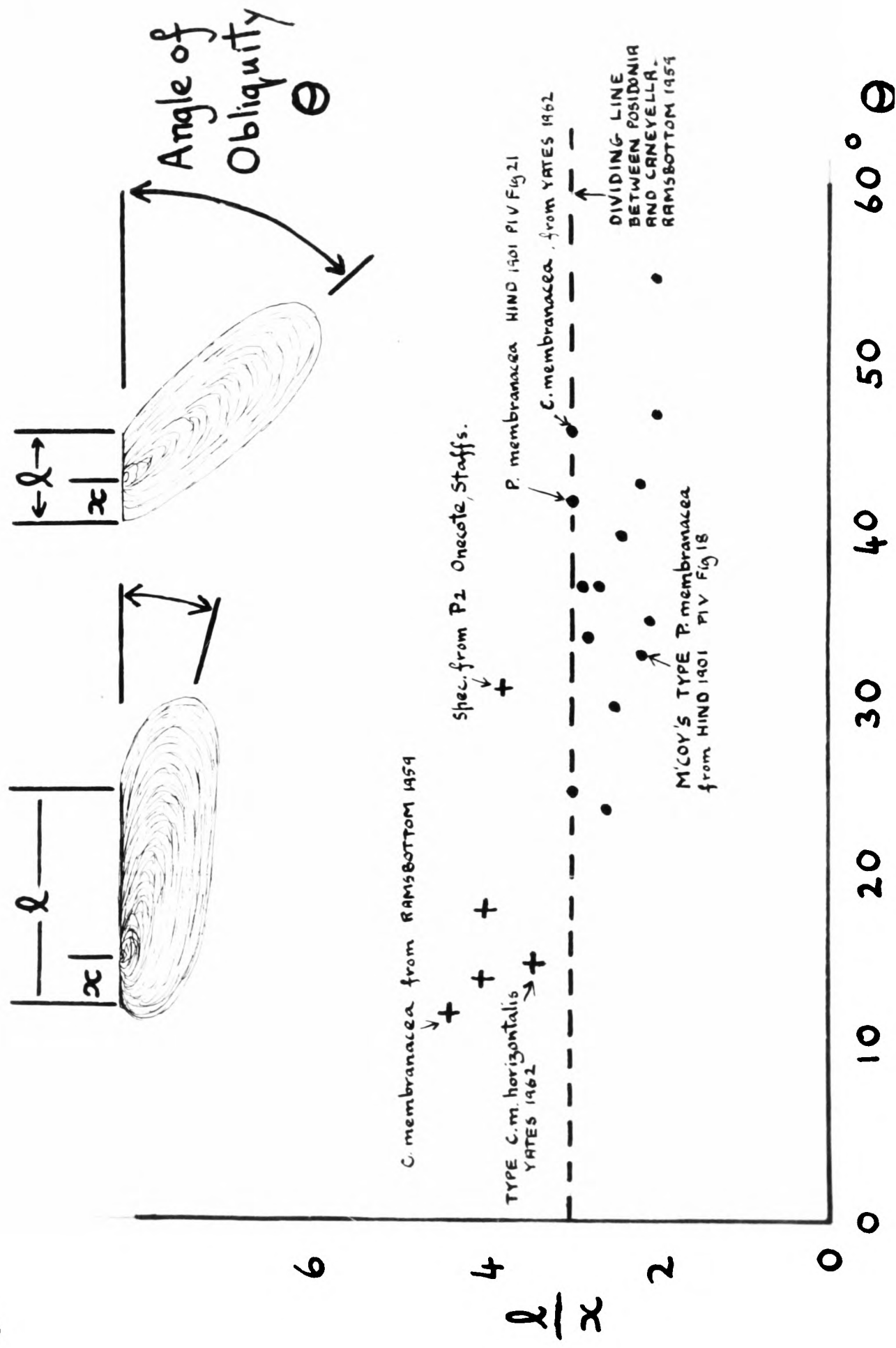
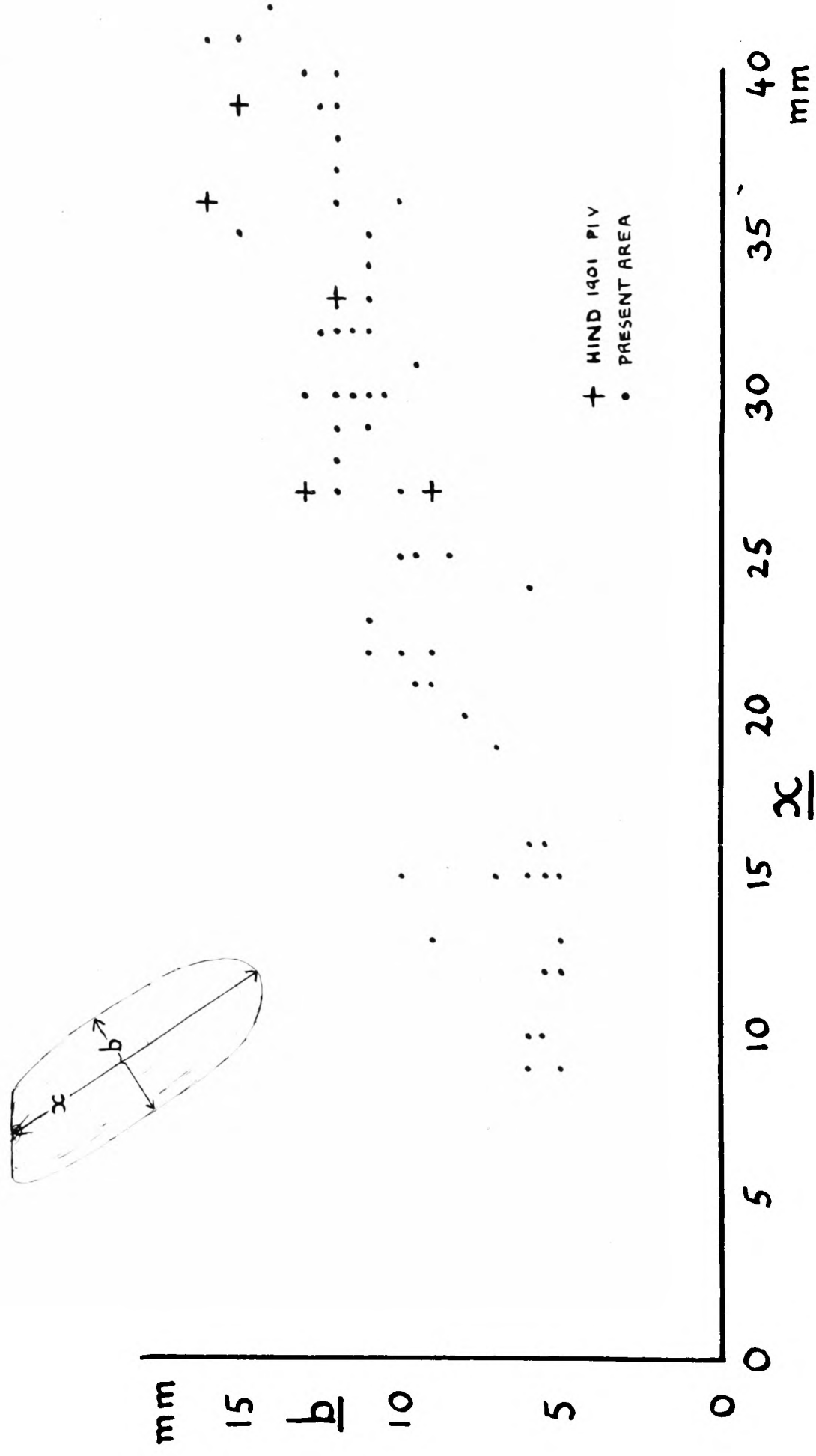


Fig 13d

PLOT of DIMENSIONS x and b for P. MEMBRANACEA



EUMORPHOCERAS GRASSINGTONENSE HORIZON.

FAUNAL LIST.

E. grassingtonense Dunham and Stubblefield Fig.1.3e 1.3g1Eumorphoceras sp. nov.I Fig.1.3g2-3Cravenoceras cf. C. cowlingsense BisatDimorphoceras sp.Cycloceras kionoforme DemanetCycloceras purvesi DemanetActinopteria persulcata (M'Coy)Obliquipecten laevis HindObliquipecten aff. costatus YatesPosidonia aff. membranacea (M'Coy) (small form)Posidonia corrugata elongata Yates Fig.1.3h1Posidonia corrugata (Etheridge)Posidoniella aff. vetusta (J de C. Sow) Fig.1.3h2

Fish remains

This is by no means an exhaustive faunal list for this horizon, there are for example several other species of nautiloids present and several distinctive forms of Posidonia. The list does however include all the species that are useful in the identification of this horizon. The most important member of the fauna is the Eumorphoceras which is identified as E. grassingtonense. The material is generally poor and always crushed.

The horizon from which the fauna is obtained

consists of from 120 to 400cm. of silty shale which passes up into calcareous siltstones. The lowest part of the faunal band contains only lamellibranchs, and the goniatites are most frequent in strong fissile shale with a regular vertical jointing just below the base of the calcareous siltstone unit. Upwards, within the calcareous siltstones lamellibranchs (P. corrugata) are present but they gradually become rarer towards the top of the calcareous siltstones and are absent in the shale-mudstones with siderite nodules that follow. Below the faunal band is 180-300cm. of unfossiliferous shale-mudstone before the first sandstone of the B. unit of protoquartzites is seen. Roughly in the middle of the shale-mudstone is a 0.5cm. pyritic clay bed (B2) which is a reliable marker horizon and whose origin is discussed in section 1.4.

DESCRIPTIONS OF SPECIES.

The specimens obtained in the present area are described and compared with the type of E. grassingtonense. An apparently new species of Eumorphoceras is also described from this horizon.

Family	GONIATITIDAE	de Haan
Subfamily	GIRTYOCERATINAE	Wedekind
Genus	EUMORPHOCERAS	Girty 1909

Eumorphoceras grassingtonense Dunham & Stubblefield Fig.1.3e

Eumorphoceras bisulcatum Girty mut. grassingtonensis

Dunham & Stubblefield 1944 p.258-260 Pl.XI Figs.4a-c

Eumorphoceras bisulcatum grassingtonensis Dunham & Stubblefield

Yates 1962 p.381 Pl. 52 Figs.4a-c

These specimens are all crushed, are often fragmentary, and the ornament is not often clearly seen. The two fragments illustrated are from the part of the shell where the ribs begin to die out. This occurs from 10 to 12mm. spiral groove diameter. At a spiral groove diameter of 10mm. the umbilicus is 4 to 4.5mm. wide (measured at outside edge). Fig.1.3f1 shows the size of the umbilicus relative to the spiral groove diameter for several crushed specimens and also for the type specimen. The typical ornament consists of ribs which arise at a node on the umbilical margin and then bifurcate over the flank. Between the bifurcating ribs lies a single non-bifurcating rib which does not have a node at the umbilical margin. The ribs show a slight forward projection for the first $\frac{2}{3}$ of their course over the flank and then bend forward to the spiral groove. The ribs number about 10 per $\frac{1}{2}$ whorl. Larger specimens show that when the ribs are barely visible over the flanks small nodes are still present at the umbilical margin. These nodes are well marked at 15mm. spiral groove diameter and are still present as tiny raised ridges at 18mm. spiral groove diameter, 23mm. overall diameter. The largest specimen

seen, with a spiral groove diameter of 20mm. and overall crushed diameter of 27mm. does not appear to have any nodes and the shell is smooth on the last whorl apart from the spiral groove which is still fairly strong. The type of rib bifurcation typical of this species does not occur on the whole of the ribbed part of the test but appears to be confined to the part just before the ribs begin to fade over the flanks. One specimen shows a possible shallow constriction but the crushing does not permit the certain recognition of constrictions.

DESCRIPTION OF TYPE SPECIMEN OF E. GRASSINGTONENSE.

E. bisulcatum Girty mut. grassingtonensis Dunham and Stubblefield. 1944. Inst. of Geological Sciences Coll. Spec.KD423. Fig.1.3g1

The specimen is solid with the shell preserved and hence no suture line is visible. The maximum diameter is c12mm. with a maximum spiral groove diameter of 9.5mm. At this diameter the umbilicus is 4.4mm. wide, and at 7.5mm. diameter the umbilicus is 3.7mm. wide.

The spiral groove becomes stronger on the whorl seen, it is weak at the smallest diameter but strong and 0.5mm. wide with a second shallow groove ventral to the main groove at the largest diameter seen. The furthest forward extension of the lingua occurs on the outer ridge of the main groove.

Constrictions are present but are not very strong. Four are present on the whorl seen but seem to be irregularly spaced and appear only faintly over the venter. They are best seen at the umbilical margin and where they cross the spiral groove.

The ribs are strong to the maximum diameter seen. The shape of the ribs is variable but most arise at the umbilical margin and start in a slightly forward direction which is maintained for $\frac{2}{3}$ of the distance to the spiral groove, in the last $\frac{1}{3}$ of their traverse over the flank they bend sharply forward to meet the spiral groove. Some ribs have a node at or near the umbilical margin and several broaden over the flank, especially those that immediately precede a constriction. No regular bifurcation is seen on the specimen but in the final part preserved a rib without an umbilical node may rise very near one with a strong node so giving the appearance of a bifurcation.

There are 13-16 ribs/ $\frac{1}{2}$ whorl. The ribs are often swollen in the mid-flank region just before the point where they swing forward to the spiral groove.

The ventral sinus is 2mm. deep at 9mm. spiral groove diameter and 0.8mm. at the smallest part of the whorl seen. The venter has an ornament of growth lines with a very faint spiral ornament which is best seen at the ventral side of the spiral groove.

Discussion.

The correspondence between the type specimen and the present material is very close in overall size and umbilical diameter. The rib type and shape is also very similar as is evident from Dunham and Stubblefield's description that:-

"There is a tendency for some of the ribs to expand so that at mid length they are thicker than their fellows, and these sometimes branch from the rib preceeding from their origin on the umbilical margin giving the appearance of a bifurcation."

They also mention that umbilical nodes are present. It is unfortunate that the type specimen is not larger as then it might show more of the part of the test with the characteristic bifurcating ribs. One bifurcation is clearly seen on the type specimen. The ribs on the type specimen appear to be stronger, but this may be due to the differences in preservation. The type specimen possesses constrictions which have not been recognised with certainty on the present material, this is not surprising considering the original weakness of the constrictions and the crushed nature of the present material. It is concluded that the present material is essentially the same as the type specimen of E. grassingtonense.

Yates (1962) described E. grassingtonense from the Slieve Anierin succession and it appears to be identical to the present material, in particular her Spec.7047, Fig.3, Plate 52 shows an identical pattern of rib bifurcation and shape and this occurs just before the ribs die over the flank which she states to be at about 12mm. diameter. Yates illustrates constrictions on Spec.7046, Fig.4, Plate 52. These occur at about 7mm. diameter and compare well with those on the type specimen.

Eumorphoceras sp. nov.I

Loc 353 Croker Hill Figs. 1.3g2; 1.3g3.

Material. Ten specimens crushed in shale.

Description.

All the specimens are small with a maximum diameter of 11mm. and a maximum spiral groove diameter of 9mm. A plot of spiral groove diameter against umbilical diameter (Fig.1.3f2) indicates that at 6mm. spiral groove diameter the umbilicus is 2.5mm. wide, and at 8mm. diameter it is 3.1mm. wide. The spiral groove starts at 5mm. diameter and is strong at 6.5mm. and has a second shallow groove on its ventral side making the ventral edge of the main groove stand out as a ridge. The ventral margin of the groove is ornamented with fine spiral ridges (in Spec.353D, 4 ridges seen. Fig.1.3g3). The spiral groove persists to the largest diameter seen.

There are 13-15 ribs/ $\frac{1}{2}$ whorl which are sharp, especially at the umbilical margin. They are generally inclined sharply forward and die out before reaching the spiral groove. At 5-6mm. diameter the ribs reduce to small sharp plications on the umbilical margin, and at 9mm. diameter only small nodes are left and these are more widely spaced at 10 per $\frac{1}{2}$ whorl (see Fig.1.3g2). The ribs occasionally group in pairs, each pair being separated by an isolated rib, this feature is reminiscent of E. grassingtonense.

There is a strong ventral sinus which is 3mm. deep at 8mm. diameter. No constrictions have been seen on the present material.

Discussion.

This species differs from E. grassingtonense in the early dying of the ribs but its rib shape and grouping and umbilical nodes and strong spiral groove are reminiscent of E. grassingtonense.

There are no other described British Eumorphoceras species which have the combination of short sharp ribs reducing at 5-6mm. diameter to umbilical nodes. This species differs from all the E. pseudobilingue group in that the ribs die out at a smaller diameter. It is closest to E. pseudobilingue C Bisat (see Yates 1962 p.380 Pl.52 Figs.1,2) in which the ribs reduce to small

plications at the umbilical edge at 15mm. overall diameter. The form of these plications at the umbilical edge is very similar to that seen in the present series, possibly indicating some relation with the earlier E. pseudobilingue group.

Subfamily	HOMOCERATINAE	Spath
Genus	CRAVENOCERAS	Bisat 1928

Cravenoceras cowlingsense Bisat 1930

C. cowlingsense Bisat 1930 p.29-30 Pl.I Figs.1-3

Cravenoceras cf. C. cowlingsense

As was the case with the specimens from the C. cf. C. malhamense band, the material is very poor and is all crushed in shale. Ventrally crushed specimens show fairly strong striae passing virtually straight across the venter. Umbilically crushed specimens, which are scarce, show the umbilicus to be wide and the umbilical edge fairly sharp. On the flanks the striae bend slightly back, but this is not so marked as in the specimen illustrated by Bisat (1930). Without solid specimens the material can only be referred to as C. cf. C. cowlingsense. In this faunal band specimens of Cravenoceras are far more numerous than those of Eumorphoceras.

NOTES ON LAMELLIBRANCHS IN E. GRASSINGTONENSE HORIZON.

Class LAMELLIBRANCHIATA Blainville 1816

Family AVICULOPECTENIDAE Etheridge Jun., emend. Newell 1937

Genus OBLIQUIPECTEN Hind 1903

Obliquipecten costatus Yates 1962

Obliquipecten costatus Yates 1962 p.401-2 Pl.59 Figs.2,3
Pl.62 Fig.5

Obliquipecten aff. costatus Yates

This species differs from O. laevis according to Yates in possessing radial ribbing. Some of the specimens obtained from this horizon do possess radial ribs, but they are not so prominent as those illustrated by Yates, hence the specimens are referred to as O. aff. costatus. Yates (1962 p.402) states that O. costatus is abundant in E1a, but also mentions that it occurs "at other horizons in E beds but never so abundantly."

Family PTERINOPECTINIDAE Newell 1937

Genus POSIDONIA Etheridge

Notes on forms present in E. grassingtonense horizon.

The specimens show great variation, and many are similar to Yates P. corrugata elongata but are larger, the maximum height being 20mm. in Yates specimens.

There is little to distinguish them from the larger P. membranacea of the C. cf. C. malhamense band except that the hinge is shorter and the angle of obliquity is greater. A typical specimen is shown in Fig.1.3h1. The corrugation of the shells is typical of P. corrugata in being stronger than that of P. membranacea.

Several specimens of a similar shape to the above but with thinner shells without the coarsely corrugate ornament of P. corrugata are also present. In shape these specimens are indistinguishable from P. membranacea but they do not exceed 20mm. in length in contrast to the normal P. membranacea which can be up to 50mm. long, they are therefore referred to as P. aff. membranacea (small form). Rare specimens of Posidoniella aff. vetusta also occur in this horizon, Fig.1.3h2.

LOCALITIES FOR THE E. GRASSINGTONENSE HORIZON.

Locality Number.	General Area.
001	} Hurdlow
008A	
020	
021	
055	Upper Shirkley
106	Pyecclough
120	
209	Blake
252	Sunnydale

Locality Number.	General Area.
353	Croker
357A	
358	
504	Elkstones
525	
581	Winkhill - Martinslow

The presence of the E. grassingtonense horizon has been proved at the above localities on a combination of sedimentological and palaeontological evidence. In all locality areas quoted above, the pyritic clay bed B2 can be found, (Section 1.4) and the sequence of lithofacies above and below the faunal band is identical. The most extensive faunal collections were made at Locs. 001, 008A, 357A and 581.

Locality Notes. References by other authors to the above localities.

LOC 581. Prentice 1951 (His Loc 120) records E. bisulcatum Girty from Loc 581, and Morris (His Loc 26) records an extensive fauna in which he includes the following goniatites:-

Eumorphoceras bisulcatum Girty

Cravenoceratoides nitidus (Phillips)

Cravenoceratoides sp. cf. Ct. fragilis Bisat

Anthracoceras sp.

Morris' identification of Ct. nitidus should place this band in E2b.

The identification of Ct. nitidus is difficult, and Morris gives little evidence to support his identification. The writer considers this identification to be incorrect since the fauna he has collected corresponds well with that from the E. grassingtonense band and at this locality includes:-

E. grassingtonense

C. cf. C. cowlingsense

Dimorphoceras sp.

O. laevis

P. corrugata

Fish remains

The specimens of Eumorphoceras are poorly preserved but one clearly shows the rib bifurcations and rib shape typical of E. grassingtonense together with weakening of the ribs at 10mm. spiral groove, 12mm. overall diameter. The Cravenoceras specimens are, as Morris mentions, mainly ventral impressions. No specimens have been seen with the bifurcations or the canted lirae typical of Ct. nitidus. The ornament is of Cravenoceras type and the specimens are described here as C. cf. C. cowlingsense.

Apart from the faunal evidence there is the occurrence of the 0.5cm. pyritic clay bed B2 below the faunal horizon which confirms its identification. Also there

is the general lithological succession of protoquartzite lithofacies below, and calcareous siltstones above the faunal band. The calcareous siltstones are overlain by shale-mudstone with siderite adding further confirmation that this is the E. grassingtonense horizon.

LOC 001. This locality is Loc 98 of Morris (Thesis 1966) and from it he records the following goniatites:-

E. bisulcatum Girty

Cravenoceras sp.

Dimorphoceras sp.

Morris suggests an E2b age for this fauna since he equates the sandstone below this horizon with the sandstone below Ct. edalensis in the Easing brook section (my Locs 461 - Ct. edalensis; 462 - protoquartzites; Morris Loc 118). He made this correlation by feature mapping which is very difficult and often misleading in this glaciated area. He recognises however that the horizon of Loc 001 is below the horizon of Ct. edalensis seen downstream (Loc 003. Morris Loc 119) and that the two horizons are separated by a thickness of sandstones and shales which he estimates as 400 feet. Thus he comes to the conclusion that there are two horizons with Ct. edalensis about 400 feet apart and that the horizon at Loc 001 is between the two. Morris thus produces a sequence of horizons for E2b which differs radically from that known in other areas.

Detailed mapping of the area by the author (see Fig.1.1e) clearly shows that the horizon at Loc 001 is below the E. erinense horizon. The horizon seen at Loc 001 is again seen at Loc 008A only 300 yards from Loc 001 (see Fig.1.1e). At Loc 008A the horizon is seen at the base of the cliff and is underlain by sandstones and overlain by calcareous siltstones which pass up into shale-mudstone with siderite. At the top of the cliff (Loc 008) the E. erinense band is exposed and yields frequent E. ferrimontanum and E. erinense. The horizon exposed at Locs 001 and 008A has yielded specimens of E. grassingtonense (see Figs.1.3e1,1.3e2) and is underlain at both localities by the pyritic clay bed B2 which is invariably found beneath the E. grassingtonense horizon in the area.

Thus the exposure found by Morris at Loc 001 is not considered to be the C. subplicatum band of Hudson (1944) and Yates (1962) as suggested by Morris, but is the E. grassingtonense horizon of Yates (1962) which equates with the E. aff. pseudobilingue horizon of Hudson and (1943, 1945, 1945a). A horizon corresponding with Yates C. subplicatum horizon is in fact present in this area and will be discussed later.

The localities in the Croker Hill area 353, 357A and 358 and Loc 055 at Upper Shirkley were recorded by Evans et al. (1968 table 2) and all were assigned to the

lower E. bisulcatum band which is in agreement with the opinion expressed here.

None of the other localities are referred to in the literature as far as is known.

COMPARISON WITH OTHER AREAS.

Slieve Anierin, Eire.

The E. grassingtonense horizon described by Yates from Ireland is undoubtedly the equivalent of the horizon with E. grassingtonense in the present area. Yates records that C. cowlingsense is much more common than Eumorphoceras and she also records the presence of Dimorphoceratids. The lamellibranch fauna on Slieve Anierin is rather poor and only P. corrugata, P. lamellosa and Chaenocardiola footii are recorded, of these, only P. corrugata is recorded from the present area. This is probably due to rather different conditions prevailing in the two areas. In Slieve Anierin the horizon is sandwiched between 'unfossiliferous shales with clay ironstones' (Yates 1962), whilst in the present area it is underlain by protoquartzites, and overlain by calcareous siltstones, some of which occur within the faunal band.

The position in the succession of Yates' E. grassingtonense horizon is also identical with that in the present area in that it is the only faunal band between

the C. cf. C. malhamense and E. erinense horizons.

Edale and Alport, Derbyshire.

Hudson and Cotton (1943 and 1945) have described the Alport and Edale boreholes and the surface exposures of the areas. Yates (1962) correlated her E. grassingtonense horizon with the E1d E. aff. pseudobilingue horizon of Hudson and Cotton. This correlation is confirmed by the present writer after examination of the Edale section where this horizon is exposed with shale-mudstone with siderite below and calcareous siltstones above. This is similar to the situation in north Staffordshire except that the protoquartzites are not present, this feature is explained in the sedimentology section. The faunal band is also underlain by the 0.5cm. pyritic clay bed B2 seen in north Staffordshire, thus providing evidence of correlation independent of the faunas. Hudson and Cotton considered the fauna had more in common with the Pendleian than the Arnsbergian, but the author follows Yates (1962) and places this band in the Arnsbergian. This also conforms with present day usage.

Ashover, Derbyshire.

Ramsbottom, Rhys and Smith (1962) in their descriptions of the faunas from the Ashover boreholes found specimens of Eumorphoceras but these were not well enough

preserved to indicate which of the E2a horizons were present. Evidence supplied by the clay beds and pyrite beds suggests that the lower E2a horizon may be present in the Tansley borehole (see Section 1.4) but this cannot be confirmed by the fauna.

Greenhow Mining Area.

Dunham and Stubblefield (1944) obtained the type specimen of E. grassingtonense from the Cockhill Limestone and this limestone also contains C. cowlingsense. Thus it has a similar goniatite fauna to the E. grassingtonense band already described, and is correlated with this horizon by Yates (1962).

Lancaster Fells.

Moseley (1954) described two horizons from the Tarnbrook Wyre Marine beds (E2a) but was uncertain as to their relative position since the exposures are at different localities. One horizon, exposed in Tarnbrook Valley (Loc 15), contains C. cowlingsense and 'E. bisulcatum', and the other horizon outcropping in a headstream of the Brennand (Loc 14) contains dominant 'E. bisulcatum'. He considers that the Tarnbrook Valley beds are the highest, but the work of Yates shows that this view is probably mistaken and she equates the Tarnbrook Valley beds with the E. grassingtonense horizon and the Brennand band with

the upper horizon. Yates states that material from the Brennand band is close to E. erinense which she finds in the higher band in Ireland and which is also present in the same band in north Staffordshire.

Explanation Figs.1.3e-h.

Fig.1.3e

1. Eumorphoceras grassingtonense Dunham and Stubblefield x5,
Spec.001A, Loc.001, Hurdlow.

2. E. grassingtonense x5, Spec.008A1, Loc.008A, Hurdlow.

3. E. grassingtonense x10, Spec.581A, Loc.581, Winkhill.

Fig.1.3f

Plot of umbilical and spiral groove diameters for E. grassingtonense (1.3f1) and Eumorphoceras sp. nov.I (1.3f2)

Fig.1.3g

1. E. grassingtonense Dunham and Stubblefield 1944. The type specimen. IGS Leeds collection No.KD423. Cockhill Limestone.
2. Eumorphoceras sp. nov.I x5, Spec.353F, Loc.353, Croker Hill.
3. Eumorphoceras sp. nov.I x5, Spec.353D, Loc.353, Croker Hill.

Fig.1.3h

1. Posidonia corrugata elongata Yates. x5, Spec.353K, Loc.353, Croker Hill.
2. Posidoniella aff. vetusta (J. de C. Sow) x2.5, Spec.001B, Loc.001, Hurdlow.

Fig 13e

1



2

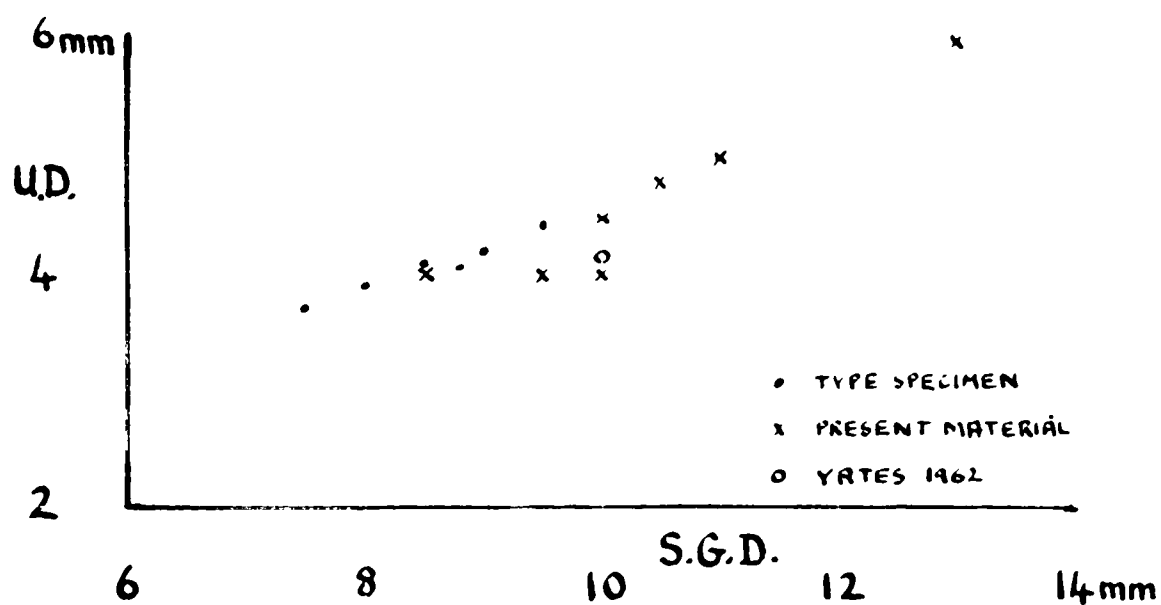


3



Figl.3fi

PLOT OF UMBILICAL AND SPIRAL GROOVE
DIAMETERS FOR E. GRASSINGTONENSE



Figl.3fii

PLOT OF UMBILICAL AND SPIRAL GROOVE
DIAMETERS FOR EUMORPHOCERAS sp. nov. I

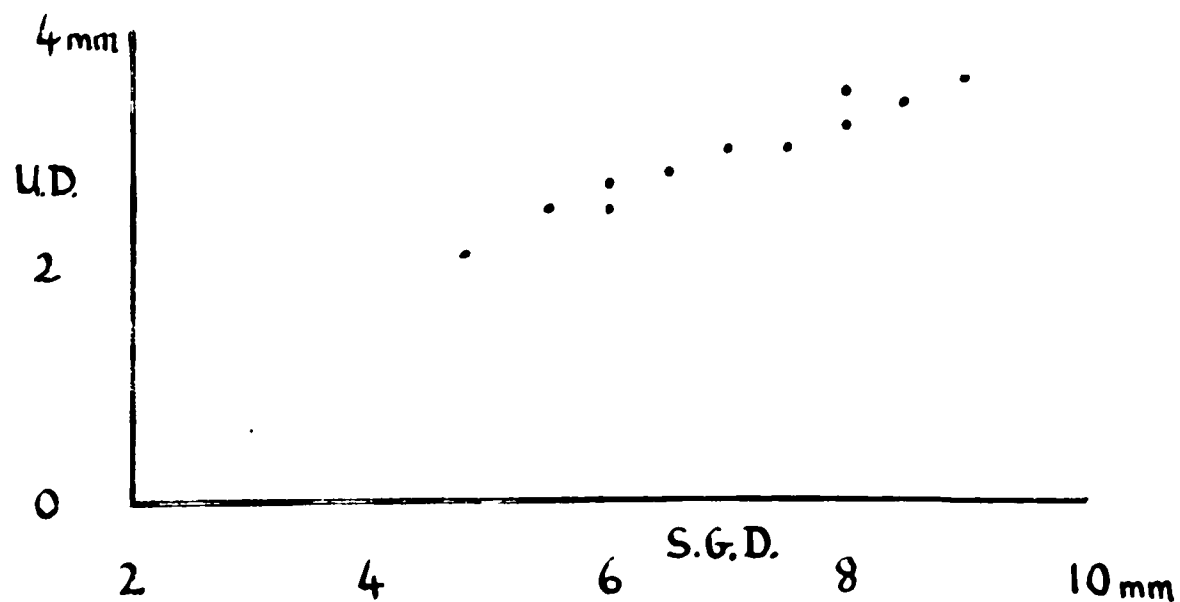
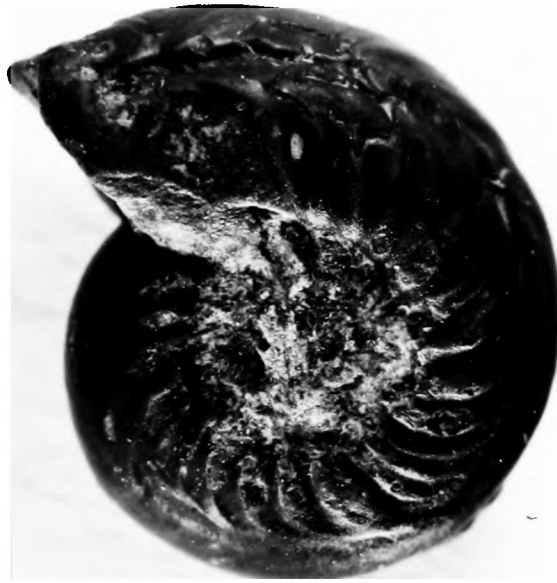


Fig 1.3g



1

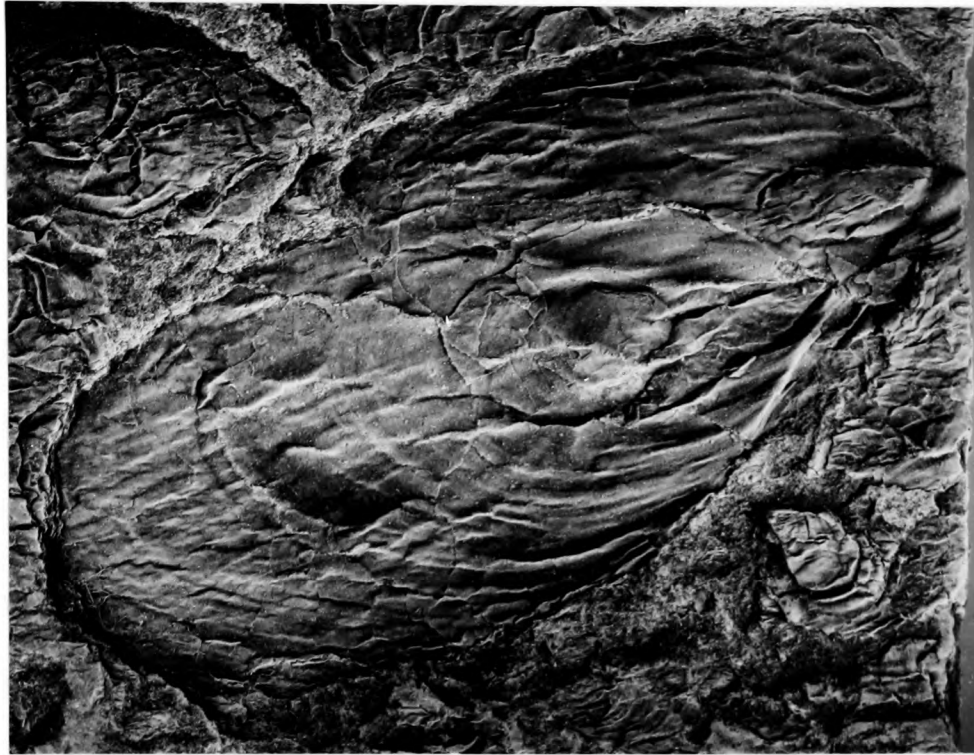
2



3



Fig 1.3h



1

2



EUMORPHOCERAS ERINENSE -EUMORPHOCERAS FERRIMONTANUM HORIZON.

This horizon is the thickest goniatite bearing horizon in the part of the succession under consideration. It is generally about 5 metres thick and goniatites occur throughout this thickness but are much more abundant at some levels than at others. Fig.1.3i shows the distribution of faunas obtained from this band at Croker Hill (Loc 351). The most useful level is near the base and contains abundant E. ferrimontanum and E. erinense.

The sediments that comprise the horizon consist of shales of black or brown colour which generally contain the fauna and thin silty beds up to about 4cm. thick which resemble the calcareous siltstone lithofacies and generally do not contain such a rich fauna.

Secondary calcification often affects the beds converting them to platy limestones in which fossils are much more difficult to find. The calcification took place after compaction of the shales and silty beds since fossils are always crushed with the exception of crinoid ossicles.

FAUNAL LIST.

<u>Eumorphoceras ferrimontanum</u>	Yates	Fig.1.3j2
<u>Eumorphoceras erinense</u>	Yates	Fig.1.3j1
<u>Eumorphoceras sp. nov.</u>		Fig.1.3j3
<u>Kazakhoceras scaliger</u>	(Schmidt)	
<u>Anthracoceras</u> - <u>Dimorphoceras</u>	sp.	
<u>Cravenoceras</u> cf. <u>C. gairense</u>	Currie	Fig.1.3j4
<u>Cycloceras purvesi</u>	Demanet	
Various nautiloids		
<u>Posidonia corrugata</u>	(Etheridge)	
<u>Posidonia corrugata elongata</u>	Yates	Fig.1.3k2
<u>Chaenocardiola footii</u>	(Baily)	
<u>Pseudamussium</u>	sp.	
<u>Dunbarella</u> aff. <u>elegans</u>	(Jackson)	Fig.1.3l2
<u>Dunbarella</u> aff. <u>papyraceus</u>	(Sow)	Fig.1.3l1
<u>Chonetes</u>	sp.	
Crinoid debris cf. <u>Blothrocrinus</u>		Fig.1.3k1
<u>Edestodus</u>	sp?	
Conodonts		

FAUNAL DESCRIPTIONS AND NOTES - THE GONIATITES.

Family	GONIATITIDAE	de Haan
Subfamily	GIRTYOCERATINAE	Wedekind
Genus	EUMORPHOCERAS	Girty 1909

There are several species of Eumorphoceras present within this horizon. Near the base of the band is a horizon with abundant E. erinense and E. ferrimontanum (Figs. 1.3j1-2). All the specimens from this level appear to conform well with Yates' descriptions. The two are best distinguished by the earlier dying of the ribs in ferrimontanum at 11mm. spiral groove diameter whereas in erinense they persist to 15mm. spiral groove diameter. (These figures computed from Yates' illustrations.) E. erinense also shows occasional rib bifurcations.

Above the level with abundant E. ferrimontanum and E. erinense specimens of Eumorphoceras are generally scarce, but the few that have been obtained appear to differ significantly from those in the lower part of the band. The ribs appear to be fewer in number and straighter with strong nodes at the umbilical margin. The best specimen obtained from the top part of this band is from Loc 523 (Fig. 1.3j3). This specimen has only 11-12 ribs per $\frac{1}{2}$ whorl even when a bifurcating rib is counted as two. The ribs are strong but may be beginning to fade at the largest diameter seen (12mm. spiral groove). Bifurcating ribs alternate with non-bifurcating ribs for most of the last whorl and the bifurcating ribs generally have a raised tubercle at the umbilical margin. The ribs are radial for $\frac{2}{3}$ of their course over the flanks, then bend strongly forward to the spiral

groove. The umbilicus is 4mm. wide at 12mm. spiral groove diameter. This specimen is comparable in many ways to E. grassingtonense but differs in its lower rib count, 11-12 as against 16 per $\frac{1}{2}$ whorl, and it differs from the E. grassingtonense material from the lower band in its stronger and straighter ribs. This is apparently an undescribed form, but it is pointless to describe it further here as only one specimen is known to the author, and the marine band in which it occurs can be easily recognised from the other fossils occurring within it.

Subfamily	HOMOCERATINAE	Spath
Genus	CRAVENOCERAS	Bisat 1928

Cravenoceras gairense Currie

Cravenoceras gairense Currie 1954 p.577 Pl.IV Figs.8-10

Cravenoceras cf. C. gairense Currie.Fig.1.3j4

Specimens of Cravenoceras are generally scarce in this horizon but two that have been found have the distinctive spiral ridges of C. gairense preserved on the umbilical edge. However, only two continuous ridges are present the third being poorly developed. The original C. gairense Currie has three spiral ridges. The striae number 3 or 4 per mm. at 6mm. from the umbilical edge as mentioned by Yates (1962) but this feature probably depends on the size of the shell. The striae

also have a slight forward bow from the umbilicus over the flanks.

This species, described by Currie (1954) from the Gair Limestone of Lanarkshire which she correlates with E2b, has been recorded by Yates (1962) from the high E2a band on Slieve Anierin. It is here recorded from the same E2a band in Staffordshire for the first time. The specimens were obtained from Loc 210 Blake and Loc 561 Waterhouses.

FAUNAL DESCRIPTIONS AND NOTES - THE LAMELLIBRANCHS.

The lamellibranchs are numerous and more varied than at lower horizons, Chaenocardiola and Dunbarella appear for the first time and Pseudamussium is often abundant. Obliquipecten, which is common in the lower E. grassingtonense horizon appears to be absent. Posidonia corrugata is abundant throughout the horizon and several subspecies may be present, but only E. corrugata elongata has been identified (Fig.1.3k2).

The specimens of Dunbarella are worthy of consideration in view of Jackson's (1927a) work on this genus.

Class	LAMELLIBRANCHIATA	Blainville 1816
Family	PTERINOPECTINIDAE	Newell 1937
Genus	DUNBARELLA	Newell 1937

Dunbarella papyraceus (Sowerby)

Pecten papyraceus Sowerby 1823. p.75 Pl.354

Pterinopecten papyraceus (Sow). Hind 1903 VolIII p.51 Pl.VII.

Figs.7,9,10,11 and 13

Pterinopecten papyraceus (Sow) Jackson 1927a Pl.I Fig.1

Pl.II Fig.5

Dunbarella aff. papyraceus (Sowerby) Fig.1.311

Specimens of Dunbarella were found in the E. erinense horizon at localities 210, 561, 351 and 501. They are generally scarce but a few well preserved specimens were obtained (Fig. 1.311). These specimens have from 71 to 75 radii on the left valve at a distance of about 20mm. from the umbo. The maximum hinge length is about 31mm. and this is equal to or less than the height of the valve. The anterior ear is separated by a shallow groove extending from the umbo to a slight indentation in the anterior margin. The ear carries about 9 radii. The The posterior border is straight where it approaches the hinge line. Intercalation of secondary and tertiary radii is seen.

In the right valve the anterior ear is divided from the main body of the shell by a deep groove and a strong narrow notch is present in the border, which can be traced back to the umbo by means of the growth lines. The posterior ear is undefined in both valves.

The umbones are approximately central on the hinge line. The ears and shoulders have well marked rhythmic ridges which fade over the central part of the shell.

Discussion.

In features of shell shape and number of radii present this form appears to have affinities with both D. papyraceus (Sow) and D. elegans (Jackson). Jackson mentions that D. papyraceus has 79 radii on the left valve and D. elegans averages 67, all other species fall below this figure. In features of the anterior ear of the left valve the specimens resemble D. papyraceus. The specimens differ from D. papyraceus in having the umbones centrally placed instead of nearer the anterior end as in D. papyraceus. The height of the valves is greater than the length in the present material, which is not the case in Jackson's figure of D. papyraceus. These specimens are best referred to as D. aff. papyraceus.

Dunbarella elegans (Jackson)

Pterinopecten elegans Jackson 1927a p.103-105 Pl.I Fig.6

Dunbarella aff. elegans (Jackson) 1927a Fig.1.312

One specimen of Dunbarella from Loc 502 differs from the above species in having an indentation of the posterior margin on the left valve and thus more closely

resembles D. elegans, and this specimen is referred to as D. aff. elegans. It has however 73 radii on the left valve a feature intermediate between D. papyraceus and D. elegans.

DISCUSSION of DUNBARELLA.

Jackson recognises in Dunbarella a trend from R1 into the Lower Coal Measures expressed in an overall increase in the number of radii on the left valve. The recognition of forms closely allied to D. papyraceus and D. elegans within the E zone of the Namurian seems to indicate that forms with many radii were in existence before the other species described by Jackson. However the forms described here are not identical with Jackson's species, differing in particular in the position of the umbo on the hinge line, and could therefore be of a different stock. More material is needed from the lower part of the Namurian before any definite conclusions can be reached.

The stratigraphic positions and number of radii of Dunbarella given by Jackson (1927a) for Dunbarella species are tabulated below together with the records of Yates (1962) and the present author.

The rest of the fauna is typical of that usually recorded from this horizon. The presence of crinoid debris is a useful feature for identifying this horizon

<u>Species.</u>	<u>No. of radii in left valve.</u>	<u>Stratigraphic position.</u>	<u>Author.</u>
<u>D. papyraceus</u>	79	Lower Coal Measures.	Jackson 1927a
<u>D. elegans</u>	67	R2-G. <u>cancellatum-G. crenulatum.</u>	Jackson 1927a
<u>D. speciosus</u>	54	R2 (<u>Mut. α</u> - Early <u>Mut. β</u>)	Jackson 1927a
<u>D. rhythmicus</u>	47	R1 (<u>R. reticulatum-R. inconstans</u>)	Jackson 1927a
<u>D. aff. D. elegans</u>	-	E2a - E2b3.	Yates 1962
<u>D. aff. D. elegans</u>	73	E2a.	Present author
<u>D. aff. D. papyraceus</u>	71-76	E2a.	Present author
<u>D. persimilis</u>	62	P1.	Jackson 1927a Eden et al. 1964
<u>D. aff. D. elegans</u>	-	P1 - P2.	Yates 1962

as crinoids do not occur in any other of the faunal bands in the area. Fig.1.3k1 shows an external mould of the calyx of a crinoid from this horizon, usually only stem fragments are found. It is possibly a Blothrocrinus (Dr. W.H.C. Ramsbottom pers. comm.). A small specimen of Chonetes sp. collected from Loc 351 Croker is the only brachiopod known to the author from this horizon in the present area.

LOCALITIES FOR E. ERINENSE-E. FERRIMONTANUM HORIZON
WITH LOCALITY NUMBERS OF OTHER AUTHORS.

008	}	Hurdlow	
009			
020			
054	-	Upper Shirkley	- - - Evans et al. 19
074		Sprink	
108		Pyecclough	
157		Oakenclough	
210	-	Blake	- - - - - Holdsworth 341-Hudson & Cotton
211			1945a p.320
251	}	Sunnydale	- - - - - Holdsworth 265
253			- - - - - Holdsworth 79
261			- - - - - Holdsworth 79
301	-		- - - - - Holdsworth 267
302	-	Upper Dove Stannery	- Holdsworth 329
305	-		- - - - - Holdsworth 327
351	-	Croker Hill	- - - - - Evans et al. 9
352	-		- - - - -

356	-	} Croker Hill	- - - - -	Evans et al. 7
362	-		- - - - -	Evans et al. 6
401		Gun Hill		
463		Easing		
501	}	Lower Elkstone		
502			- - - - -	Morris 7?
523		Lower Elkstone - Heath House		
561	-	Waterhouses	- - - - -	Morris 56
622	-	Waterhouses Rail Cutting	- -	Morris 33

Holdsworth localities Thesis 1963

Morris localities Thesis 1966

Evans et al. localities 1968 Geol. Surv. G.B. Mem.110

CORRELATION WITH PREVIOUS AUTHORS WORK WITHIN THE PRESENT AREA.

Holdsworth (Thesis 1963) considered his localities 261, 265, 267, 329 and 79 all to be the E. bisulcatum s.s. (E. erinense) horizon and the present writer agrees with this interpretation. Loc 210 (Holdsworth Loc 341) was considered by him to represent a horizon lower than the E. erinense horizon since he found a specimen of E. aff. grassingtonense within it (see Holdsworth 1963, p.17). This specimen is indeed very similar to grassingtonense and identical with the grassingtonense material from the lower horizon. The present author has found E. ferrimontanum at this locality together with

C cf. C. gairense both of which are indicative of E. erinense horizon. The specimens of O. laevis from this exposure obtained by Holdsworth and identified by Ramsbottom are considered by the author to be Pseudamussium sp. Dunbarella is of frequent occurrence at this locality and in this area seems to be confined to the E. erinense horizon in this part of the succession.

Mapping of the immediate area (Fig.1.1d) shows clearly that this exposure is in the E. erinense horizon and also shows the E. grassingtonense horizon at Loc 209 upstream of 210 to be the lower Eumorphoceras faunal horizon. It can only be concluded that E. grassingtonense has an extended range and occurs rarely in the E. erinense horizon.

Of the three localities in this list mentioned by Morris (1966) two are correlated by him with E. bisulcatum s.s. (E. erinense). The third, Morris' Loc 7, he considered to be in E1 on the basis of some poorly preserved material. The writer considers that his Loc 7 is probably Loc 502 but it may represent Loc 504, in either case it is within E2a.

The localities in the Croker Hill area and at Upper Shirkley are also recorded by Evans et al. (1968 Table 2) and are all assigned to the upper E. bisulcatum band, thus agreeing with the interpretation given here.

CORRELATION WITH OTHER AREAS.

This horizon is very extensive and has been recognised in Edale (Hudson and Cotton 1945) where it contains a similar fauna including Crinoid ossicles, 'E. bisulcatum', K. (Neodimorphoceras) scaliger, Cravenoceras sp., Chaenocardiola, Pseudamussium and P. corrugata. The only difference of interest is the record of Posidoniella which is not found in the present area.

In the Alport borehole (Hudson and Cotton 1943) the beds passed through from 399-409 feet with Crinoid ossicles, 'E. bisulcatum', and N. scaliger undoubtedly represent this horizon. Cravenoceras is not recorded from this level in the Alport bore.

Yates' higher E. bisulcatum band on Slieve Anierin was the source of her subspecies erinense and ferrimontanum and she also records C. cf. C. gairense and K. scaliger. This goniatite assemblage is identical with that from the present area with the exception of the single E. grassingtonense individual recorded by Holdsworth (Thesis 1963). Yates also records the presence of Chaenocardiola footii, Dunbarella and Pseudamussium.

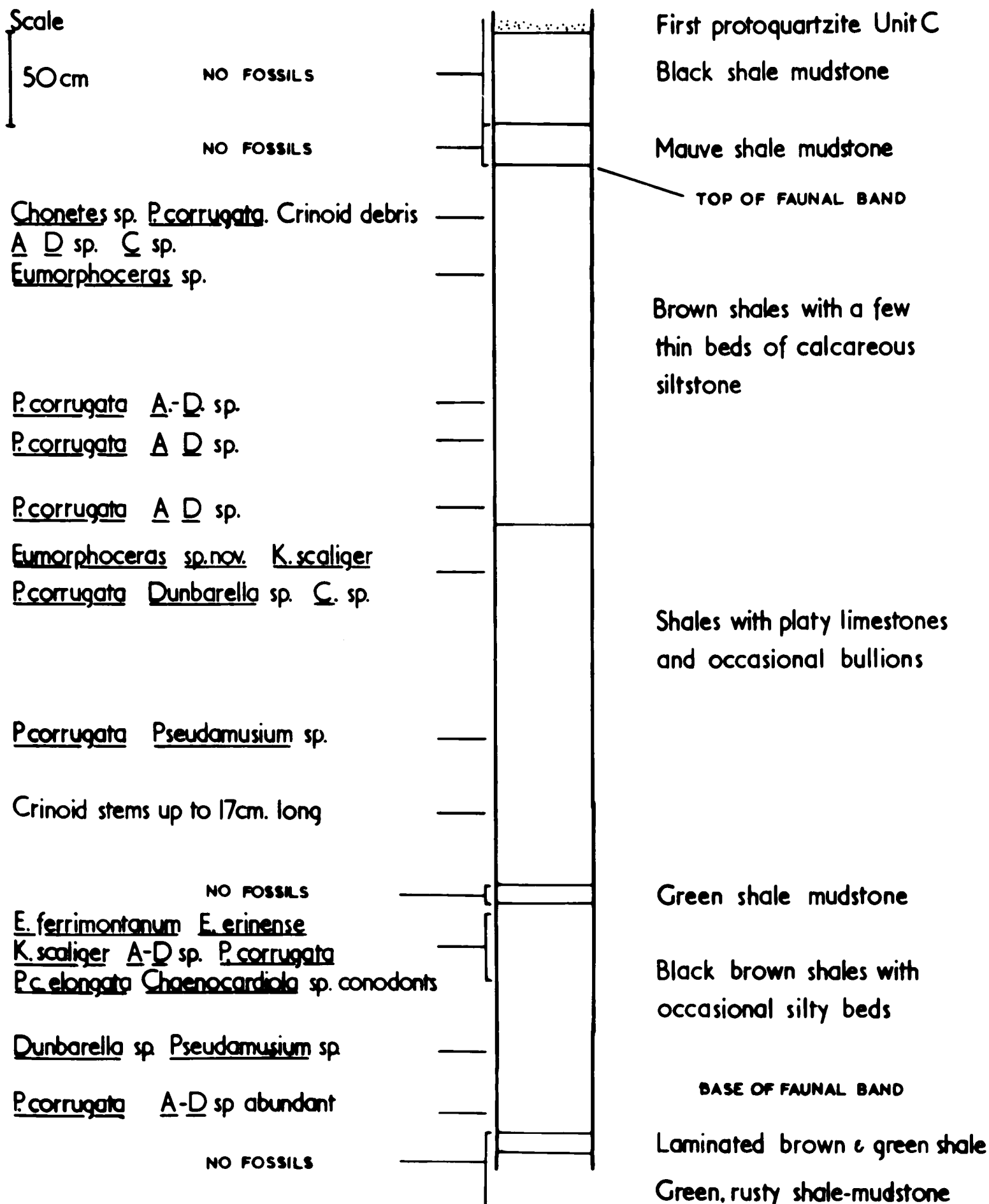
Definite correlation cannot be made with the Ashover boreholes since the specimens of Eumorphoceras obtained from the boreholes could not be identified with sufficient certainty.

The equivalent beds in the north of England have not been examined by the writer, but Yates (1962) correlates this horizon with the Warley Wise (Bray 1927), the Edge, and the Weston Marine Bands (Stephens et al. 1953). Yates also correlates the Brennand band of Moseley (1954 p.429-430) with the E. erinense horizon.

Bouckaert and Higgins (1963) also record two levels with 'E. bisulcatum' from the Dinant Basin. They record 'E. bisulcatum bisulcatum' from both horizons and also E. ferrimontanum. From the lower horizon E. grassingtonense is recorded and also C. cowlingsense. Thus the situation is similar to that prevailing in England and Ireland with the exception that E. ferrimontanum is recorded from the lower band.

Fig 13i

LOG OF *E. erinense* *E. ferrimontanum* BAND Loc 351



Explanation Figs.1.3 j-l.

Fig.1.3j.

1. Eumorphoceras erinense Yates. x5, Spec.1F214, Loc.352,
Croker Hill.
2. Eumorphoceras ferrimontanum Yates. x2.5, Spec.1F215,
Loc.352, Croker Hill.
3. Eumorphoceras sp. x5, Spec.523A, Loc.523, Elkstones.
4. Cravenoceras cf. C. gairense Currie. x5, Spec.210A,
Loc.210, Blake.

Fig.1.3k.

1. Crinoid calyx, Elothrocrinus sp. x5, Loc.352, Croker Hill.
Specimen at Inst. Geol. Sciences. Leeds.
2. Posidonia corrugata elongata Yates. x5, Spec.210B,
Loc.210, Blake.

Fig.1.3l.

1. Dunbarella aff. papyraceus (Sow). x2.5, Spec.210C,
Loc.210, Blake.
2. Dunbarella aff. elegans (Jackson). x2.5, Spec.501A,
Loc.501, Elkstones.

All specimens illustrated in Figs.1.3j-l are from the E2a2

E. erinense - E. ferrimontanum horizon.

Fig 1.3j



1



2

3



4



Fig 13k



1



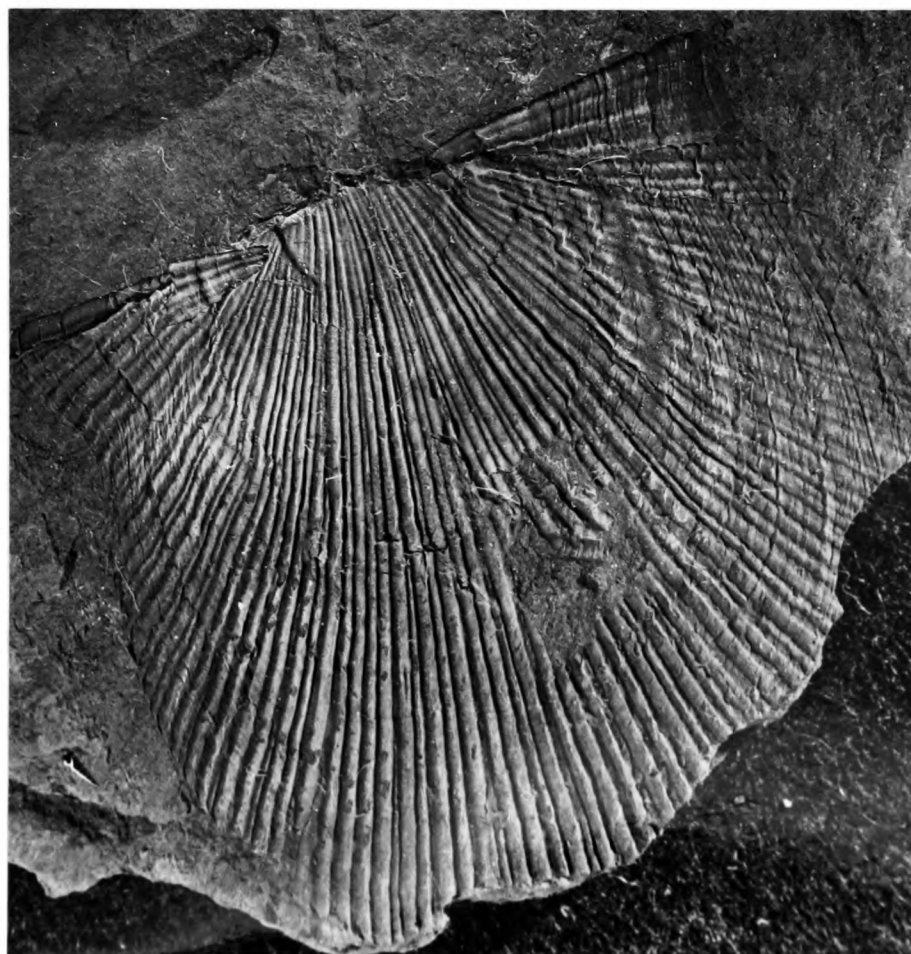
2

Fig 132

1



2



THE FAUNAL HORIZONS ABOVE UNIT C OF THE PROTOQUARTZITES.

Several distinctive faunal and lithological horizons occur within the shales and shale-mudstones that lie on top of unit C of the protoquartzites. Fig.1.3m shows the relative positions and thicknesses of these horizons up to the Ct. edalensis horizon.

The horizons are described in order starting with the L. longirostris horizon.

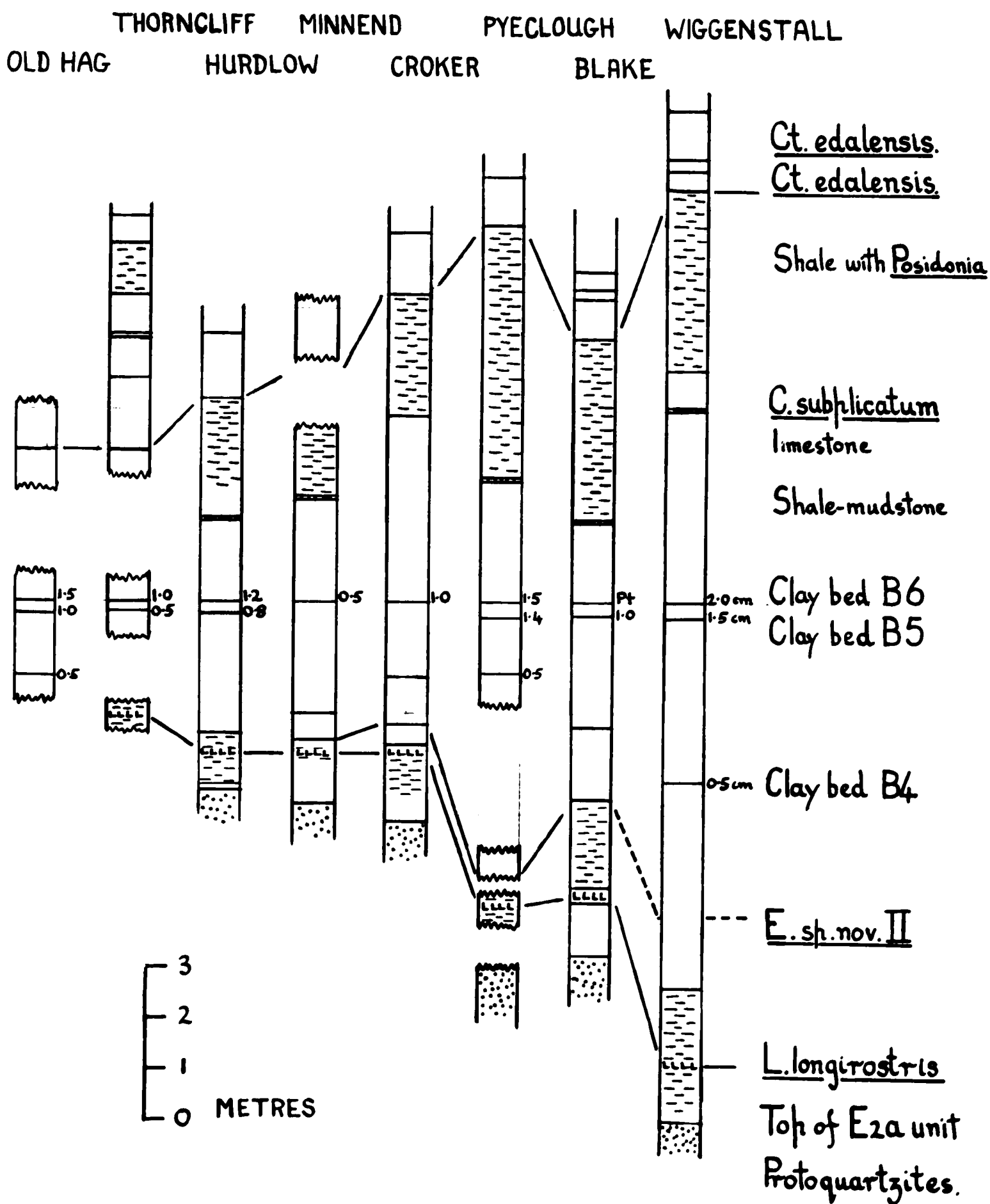
1. THE LEIOPTERIA LONGIROSTRIS HORIZON.

This is the first faunal horizon found above the C. unit of protoquartzites. L. longirostris occurs in the first dark shale present above the sandstones and specimens may be very abundant, but generally only within about 10cm. of shale. The specimens usually occur crowded on individual bedding planes with no individuals within the intervening shales, occasional seams with P. corrugata may also occur, but P. corrugata is usually absent from bedding planes that have L. longirostris.

L. longirostris has not been found at any other level within the part of the succession being studied by the author, but it has been recorded from other levels in the Namurian, e.g. Hudson and Cotton (1945) record this species from E2c in Edale.

Fig 13m

STRATIGRAPHY of the INTERVAL from L. LONGIROSTRIS
to CT. EDALENSIS.



In the present area the horizon has been found at the following localities:-

002	Hurdlow
101	Pyecclough
201	Blake
203	
382	Higher Minnend
447	Wiggenstall
461	(30 yards upstream) Easing

At all the above localities this horizon can be demonstrated to lie immediately above the C. unit of protoquartzites.

Holdsworth (Thesis 1963) obtained L. longirostris from an identical horizon in the Upper Dove near Stannery Farm, and also (pers. comm.) from the stream above Thorncliff village, near Loc 422, which again is immediately above the C unit of protoquartzites.

CORRELATION WITH OTHER AREAS.

Edale, Derbyshire.

L. longirostris has been found by the author in an identical position in the Edale area near Loc 112 of Hudson and Cotton (1945). It is found in the bank of a small tributary about 12 yards from the point where it enters the main stream. In Edale the L. longirostris horizon also appears to be the first faunal

horizon above the E. erinense band, but here the intervening unfossiliferous beds consist of shale-mudstone with siderite. The present author's stratigraphic section of the Edale exposures is shown in Fig. 1.1b.

Ashover, Derbyshire.

In all three of the Ashover boreholes (Ramsbottom et al. 1962) a horizon with L. longirostris is recorded near the top of E2a below the double horizon of pale green mudstone beds which is taken as the top of E2a in the boreholes. The fixing of the boundary at this level in the Ashover boreholes was quite arbitrary, but on the present work this horizon does divide off conveniently the E2a and E2b faunas. Both in the present area and in Edale the equivalents of these pale green mudstone beds can be found in clay beds B5 and B6 as will be shown in section 1.4. It is probable that the L. longirostris horizons in the boreholes correlate with those in the present area and in Edale.

This horizon has not been recorded in the literature on any other areas, but it is possible that it has been overlooked. However bottom conditions were probably unsuitable for L. longirostris in some areas as seems likely in the Lancaster Fells area where Moseley (1954) records that the Ct. edalensis horizon rests directly

on grits, thus grit deposition was probably taking place in this area while L. longirostris was making its sudden, but short lived, migration in the Southern Pennines.

2. EUMORPHOCERAS sp. nov. II HORIZON.

This horizon appears to be discontinuous since it has not been found at all localities where it would be expected to occur. As can be seen from the stratigraphic chart (Fig. 1.3m) this horizon lies just above the L. longirostris horizon. The fauna occurs in a black shale often with rusty surfaces and the fauna is always crushed.

FAUNAL LIST.

Eumorphoceras sp. nov. II

Figs. 1.3n, 1.3o

Anthracoceras - Dimorphoceras

Cravenoceras sp.?

Orthocone nautiloids

Posidoniella variabilis Hind

Posidonia corrugata (Etheridge)

FAUNAL DESCRIPTIONS.

Family	GONIATITIDAE	de Haan
Subfamily	GIRTYOCERATINAE	Wedekind
Genus	EUMORPHOCERAS	Girty

Eumorphoceras sp. nov. II Figs. 1.3n, 1.3o

Material. About 20 specimens crushed in shale.

Localities. Specimens of the goniatite have been found at three localities in the present area, 101, 201

and 354. The specimens from the first two localities are rather poor, but numerous specimens have been found at Loc 354 Croker Hill.

Description.

Two variants of this species appear to be present in this horizon. Up to 8mm. spiral groove diameter all specimens have similar features. At 8mm. spiral groove diameter the umbilicus measures about 3mm. and at 4mm. diameter it is 2mm. wide. A spiral groove is present at 5mm. diameter and persists to the largest diameters seen on all specimens.

Ribs are present on the second whorl where there are about 12 per $\frac{1}{2}$ whorl. With increase in diameter ribs become more numerous until there are 17-20 per $\frac{1}{2}$ whorl at 7-8mm. diameter. The ribs arise within the umbilicus and are radial or slightly forwardly directed for the first part of their course over the flanks. At about half way from the umbilicus to the spiral groove they bend evenly forward to meet the spiral groove.

Constrictions were apparent on some specimens which appear as deep grooves between two sharp ribs. In some cases there appear to be 4 ribs between constrictions but this feature does not appear to be constant. On specimens greater than 7-8mm. diameter two forms

can be distinguished.

Dimorph A.

In Dimorph A the ribs begin to die out at 8mm. diameter. This is best seen on Spec.354Q, Fig.1.3n1 in which the ribs die over the flanks at 8mm. diameter and nodes begin to develop on the umbilical margin, and only growth lines are present over the flanks. The spiral groove is still strong and has a subsidiary groove ventral to it. In the last part of the shell there are 13-14 nodes per $\frac{1}{2}$ whorl before the shell becomes smooth at about 11mm. diameter.

Dimorph B.

In this dimorph ribs are still present at 11mm. diameter but are weak in comparison with those on the earlier part of the shell. The ribs are strongest near the umbilicus but true nodes are not present. The spiral groove is still strong and there are about 17 ribs per $\frac{1}{2}$ whorl. An occasional rib bifurcation may occur. This form is typified by Spec.354P, Fig.1.3n2.

A plot is illustrated of umbilical diameter relative to spiral groove diameter (Fig.1.3p). This is bound to be rather inaccurate considering the crushed nature of the material but nevertheless shows the general relation of the two parameters. The large spread at the upper end may be due to differences in umbilical size of the

two forms after 8mm. diameter. It appears that Dimorph A has the smaller umbilicus.

DISCUSSION.

These two forms of Eumorphoceras are distinct from all others that have been described. E. leitricense Yates (1962) is similar in being of small size, having early fading of the ribs and also constrictions in the ribbed part. However it appears from Yates' figures that there are only 12-13 ribs per $\frac{1}{2}$ whorl.

E. stubblefieldi Moore has ribbing that dies at 8mm. diameter and also constrictions, but the spiral groove does not appear until 8mm. diameter and the ribs are not as numerous or as strong as in the present material.

E. bisulcatum var. varicata Schmidt differs in having short ribs and only 13-15 ribs per $\frac{1}{2}$ whorl, also Schmidt (1934) shows it to be strongly ribbed at 9mm. diameter.

E. bisulcatum mut. β Schmidt is characterised by pairing of the ribs. This does not occur in the present material, and moreover the rib count of E. bisulcatum mut. β is about 14 per $\frac{1}{2}$ whorl and the ribs persist to at least 10mm. diameter.

The combination of small size, up to 20 strong ribs per $\frac{1}{2}$ whorl at 7-8mm. diameter, and the presence

of constrictions immediately distinguishes both these forms from any others previously described. It is considered best to regard these forms as being two dimorphs of the same species since they are identical in their early stages and both occur at the same horizon and appear to be confined to that horizon.

NOTE ON DIMORPHISM IN EUMORPHOCERAS.

It is possible that the two forms of Eumorphoceras sp. nov. II represent sexual dimorphs of the same species. Sexual dimorphism is strong in some recent cephalopods, e.g. Argonauta argo, and many authors have described sexual dimorphs in ammonite genera (Makowski 1962, Palframan 1966). The two dimorphs differ in size, the male being smaller than the female at maturity and often having an aperture modified by lappets. The female has more whorls than the male at maturity, there being a definite morphological hiatus between the two forms. The initial stages of formation are identical or nearly so. Differences in ornamentation are often apparent in the later stages of development.

Makowski describes dimorphism of a similar kind in the Devonian goniatite genera Manticoceras, Tornoceras and Cheiloceras. Demanet (1943) describes dimorphism in Gastrioceras where one form is characterised by a wide umbilicus and flattened whorl section and the

other form has a narrow umbilicus and higher whorl section. The form with the wide umbilicus is interpreted by him as the female, but he gives no indication of the relative sizes of the forms at maturity.

In the present material differences in size at maturity cannot be demonstrated, but differences in ornamentation in the later stages are seen. In the ammonites the forms with the ornamentation modified at the smallest diameter is the male form, e.g. Creniceras renggeri (Oppel) Palframan 1966. Therefore in Eumorphoceras sp. nov. II dimorph A is very tentatively regarded as the male form, and dimorph B, the form with the more conservative and persistent ornament, as the female.

Another horizon that yields specimens of Eumorphoceras is the E. erinense horizon. Near the base of this horizon the species E. erinense and E. ferrimontanum occur together, and in the present area and in Ireland are confined to the same marine band, although Yates claims that erinense occurs slightly above ferrimontanum. The difference between these species is the earlier dying of the ribs in ferrimontanum (11mm. spiral groove diameter) leaving only short raised ridges on the umbilical margin, whereas in erinense ribs persist to 15mm. spiral groove diameter, and occasional bifurcations occur. Thus the differences are the same as in the two dimorphs of Eumorphoceras sp. nov. II.

A third pair of Eumorphoceras species with similar differences in ornamentation are E. grassingtonense and Eumorphoceras sp. nov.I from the E. grassingtonense horizon.

Thus three pairs of Eumorphoceras can be found in the three Eumorphoceras bearing horizons in E2a all with similar ornamentation differences. This gives some weight to the theory of sexual dimorphism, but without solid specimens of mature shells no absolute size differences can be demonstrated.

FAUNAL NOTES FOR EUMORPHOCERAS SP. NOV.II HORIZON.

The rest of the fauna from this high E2a horizon is rather poor in number of species. Only one doubtful fragment of Cravenoceras was found. P. corrugata is abundant and generally of small size. Posidoniella variabilis appears for the first time in the succession. This species is often recorded from lower in the succession (Hudson and Cotton 1945, p.10-12, E1b-E2a), but in the present area this is the lowest horizon at which definite examples of Posidoniella variabilis have been found. The presence of dominant Eumorphoceras and the lack of Cravenoceratoides places this horizon in the top of E2a. Thus in the present area there are three horizons in E2a with Eumorphoceras species.

LOCALITIES IN THE PRESENT AREA.

101	Pyecclough
201	Blake
354	Croker
380	Minnend (no <u>Eumorphoceras</u> found)

COMPARISON WITH OTHER AREAS.

The only other record of a Eumorphoceras fauna from about this level is that of 'E. bisulcatum' fragments at 367-371 feet in the Alport borehole, which is in the same position between the E. erinense horizon and the Ct. edalensis horizon. The specimens have been examined by the author but they are not well enough preserved to permit comparison with Eumorphoceras sp. nov.II. The only Eumorphoceras specimen preserved Zh.1023 shows only a portion of a spiral groove. A Cravenoceratoides fragment Zh1022 which was only one foot above in the borehole may indicate that this horizon is slightly higher than that of Eumorphoceras sp. nov.II.

The author has found a similar horizon just above the L. longirostris horizon in Edale with Posidoniella variabilis, P. corrugata and Anthracoceras or Dimorphoceras but was unable to find any specimens of Eumorphoceras. This fauna is identical with that found at Loc 380 (Minnend) where Eumorphoceras was also absent at this level.

Explanation Figs.1.3 n-p.

Fig.1.3n.

1. Eumorphoceras sp. nov.II Dimorph A. x10, Spec.354Q,
Loc.354, Croker Hill.
2. Eumorphoceras sp. nov.II Dimorph B. x10, Spec.354P,
Loc.354, Croker Hill.

Fig.1.3o.

1. Eumorphoceras sp. nov.II. x10, Spec.354R, Loc.354,
Croker Hill.
2. Eumorphoceras sp. nov.II. x5, Spec.354E, Loc.354,
Croker Hill.
3. Eumorphoceras sp. nov.II. x10, Spec.354B, Loc.354,
Croker Hill.
4. Eumorphoceras sp. nov.II. x5, Spec.354A, Loc.354,
Croker Hill.

Fig.1.3p.

Plot of umbilical and spiral groove diameters for
Eumorphoceras sp. nov.II.

Fig 1.3n



1
2



Fig 130



1



2

3

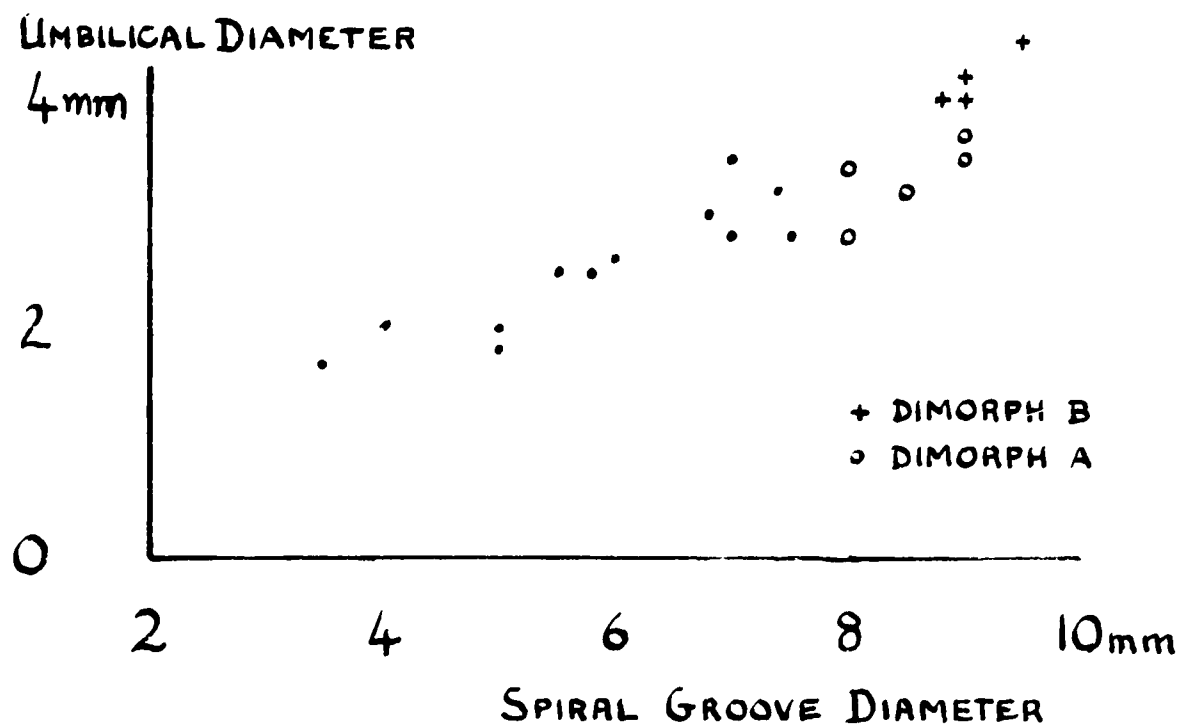


4



Figl.3p

RELATION BETWEEN UMBILICAL AND SPIRAL
GROOVE DIAMETERS FOR EUMORPHOCERAS sp. nov. II



CRAVENOCERAS cf. C. SUBPLICATUM LIMESTONE HORIZON.

The only good exposures of this horizon are in the Blake Brook at Locs. 202 and 204. At these localities the horizon consists of a continuous semi-bullion limestone band varying from 4 to 15cm. in thickness. The sediment forming the limestone has a weak parallel lamination and the goniatites are preserved as solids indicating calcification of the horizon before compaction of the sediment. This can also be demonstrated by tracing individual laminae out of the limestone into the compacted shale where goniatites are always crushed. At both the above localities the horizon occurs within calcareous shales.

At Loc. 201 only 100 yards from Loc. 202 the same horizon is represented by a sticky clay and the enclosing shales are not calcareous. This is the usual situation in the area and it appears that the shales and the limestone have been decalcified at most localities in recent times.

Holdsworth (Thesis 1963, p.112-113) describes and figures the goniatites from this limestone and identifies them as C. cf. C. subplicatum. No other species have been found in the limestone by the author but Holdsworth (1963 p.113) records a single 'Dimorphoceras' sp.

The fauna in the limestone is considered to be the lowest fauna of E2b1. This conforms with the division

made in the Ashover boreholes based on correlation of the clay beds present in this part of the succession (see section 1.4). Hudson (1945a p.2,4) records a C. subplicatum fauna below Ct. bisati which would seem to correlate with this horizon.

The original C. subplicatum figured by Bisat (1924) is very similar to the present material and came from "Birstwith Beck near Hampsthwaite at an horizon about 10 feet below Cravenoceras aff. edalense."

Holdsworth (Thesis 1963) was uncertain about the position of this fauna due to the initial identification of the Ct. edalensis band of Loc 213 as Ct. stellarum (Holdsworth Thesis 1963 Loc 277), an identification which he suggests is incorrect in the addenda to his thesis. The present author has found the Ct. edalensis band in the cliff at Loc 202 (Holdsworth Thesis 1963 Loc 271) thus confirming that the C. cf. C. subplicatum limestone lies below the Ct. edalensis horizon.

Holdsworth also records this horizon from near Stannery Farm (Holdsworth Thesis 1963 Loc 266) in the Upper Dove where it occurs six feet below a horizon with rare Ct. cf. bisati and C. cf. subplicatum.

The fauna from Loc 001 regarded by Morris (Thesis 1966) to belong to this horizon has already been discussed, and placed in the E. grassingtonense horizon.

CRAVENOCERATOIDES EDALENSIS HORIZON.

This horizon consists of shale of black or grey colour often weathering to a buff colour. The fossils are usually preserved crushed in shale but occasional bullions occur with a fauna preserved uncrushed. It can be seen from the stratigraphic chart (Fig.1.3m) that there are often two goniatite bearing horizons at this level, and as no marked differences have been found in the faunas from the two levels they are here described together.

The two horizons, when present are usually separated by about 20cm. of shale-mudstone, but at Easing (Loc 461) in the Thorncliff area 260cm. separates two horizons with Ct. edalensis. Rare examples of Ct. aff. bisati occur in both horizons. The occurrence at Easing is unusual in that the specimens are often crowded on single bedding planes and many appear to be juvenile whereas normally the specimens are scattered throughout about a metre of shale.

FAUNAL LIST.

Cravenoceratoides edalensis (Bisat) Fig.1.3q1

Cravenoceratoides bisati Hudson Fig.1.3q2

Anthracoceras sp. or Dimorphoceras sp.

P. corrugata (Etheridge)

P. corrugata aff. elongata Yates

Pseudamussium sp.

Productid fragment

FAUNAL DESCRIPTIONS.

Subfamily	HOMOCERATINAE	Spath
Genus	CRAVENOCERATOIDES	Hudson 1941

Cravenoceratoides edalensis (Bisat)

Cravenoceras edalense Bisat 1928 p.132 Pl.VI & Pl.VIa
Figs.4 & 4a.

Cravenoceratoides lirifer Hudson 1946 p.380-385 Pl.XXI
Figs.1-3,5-7.

Cravenoceratoides edalense (Bisat) Yates 1962 p.392 Pl.56
Fig.5 Pl.57 Fig.2.

The specimens have been obtained from many localities and show the typical strong tented lirae often with a single bifurcation near the umbilical margin. The lirae have a slight forward bow over the broad venter. The lirae are still strong at the largest diameters but are rather irregularly spaced and a weak lira may be present between two strong ones. Fig.1.3q1 is typical of the crushed material from this horizon.

Solid specimens of Ct. edalensis from Loc 003 Hurdlow show the median cross section illustrated for Ct. lirifer by Hudson (1946). Yates (1962) expresses

the opinion that these two species are identical and this is also considered to be the case by the present writer, after examination of Hudson's type material of Ct. lirifer.

Cravenoceratoides bisati Hudson

Cravenoceratoides bisati Hudson 1946 p.376-80 Pl.21

Figs.4,10 text Fig.1a

Some specimens of Cravenoceratoides found associated with Ct. edalensis show lirae with a double bifurcation interspersed with others with a single bifurcation or no bifurcation at all. This feature of irregular bifurcation is typical of Ct. bisati and is the main feature used to differentiate this species from Ct. edalensis (see Hudson 1946). These specimens are rarer than Ct. edalensis.

Solid specimens of Ct. bisati were obtained from Loc. 060 showing the typical ornament and also the flattened whorl section and wide umbilicus mentioned by Hudson (1946) as typical of this species. The cross section of Ct. edalensis shows a higher whorl section and a smaller umbilicus. Fig.1.3q2 shows a typical solid specimen of Ct. bisati.

Discussion.

The present work confirms that these two forms are distinct and that they occur together in the same horizon. Thus this horizon, like the Eumorphoceras horizons, contains two closely allied forms of the same genus.

It seems possible that these two species are in fact sexual dimorphs of the same species. The shapes correspond well with the shapes observed in co-habiting pairs of Gastrioceras (Demanet 1943) in that one form has a wider umbilicus and more compressed whorl section than the other. Demanet interpreted the form with the wide umbilicus as the female and that with the more narrow umbilicus as the male. On this basis in the present pair Ct. bisati is probably the female and Ct. edalensis the male.

The relative abundance of the forms has not been estimated due to the difficulty in obtaining enough solid specimens that can be determined with certainty, but Ct. edalensis would appear to be abundant and Ct. bisati scarce. The relative sizes of the two forms at maturity are not known, thus they cannot be assigned to macro- or micro-conch forms as is usually done in the case of ammonite dimorphs.

All the remarks expressed on dimorphism exhibited by the goniatites from the four major goniatite bearing horizons in the succession studied must be regarded

as tentative. Until solid specimens of mature shells are collected in fairly large numbers, (probably an impossible task in the present area) none of the required evidence of mature shell size, aperture shape, or whorl cross section is available in sufficient quantity.

LAMELLIBRANCHS.

Family	PTERINOPECTINIDAE	Newell 1937
Genus	POSIDONIA	Bronn 1828

Posidonia corrugata elongata Yates

Posidonia corrugata elongata Yates 1962 p.396-7 Pl.60 Fig.1

Posidonia corrugata aff. elongata Yates

Specimens of a P. corrugata variant from this horizon have a dimension of up to 16mm. measured from the umbo to the ventro-lateral border, with an anterior-posterior dimension of 9mm. In general shape and in the oblique nature of the anterior margin the specimens resemble P. corrugata elongata. The corrugated ornament is strongest near the umbo and tends to fade on the posterior and ventral margins. Faint radial costae are sometimes visible on the ventral part of the shell, a feature tentatively referred to by Yates (1962 p.396). Considerable variation seems to exist within the specimens from this horizon and the author here identifies all the

elongate varieties of P. corrugata whether faintly costate or non-costate as P. corrugata aff. elongata.

GENERAL FAUNAL NOTES.

The rest of the fauna is unremarkable except for the presence of a single Productid at Loc. 107, the only one found in this part of the succession, although they do occur in the E2b3 Chert horizon within the basin.

LOCALITIES IN THE PRESENT AREA.

- | | | |
|------|-----------------|---|
| 002 | Hurdlow | This is Hudson's <u>Ct. lirifer</u> locality
(Hudson 1946 p.381) Morris (Thesis
1966 Loc.119) |
| 003 | Hurdlow | |
| 060 | Dalehouse Wood. | Evans et al. 76 (1968) |
| 107 | Pyecloough | Holdsworth 260 (Thesis 1963) |
| 159 | Oakencloough | Holdsworth 226 (Thesis 1963) |
| 201 | } Blake Brook | |
| 202 | | |
| 212 | | |
| 213 | | - - - - - Holdsworth 277 (Thesis 1963) |
| 214 | | |
| 303 | Upper Dove | Holdsworth 245 (Thesis 1963) |
| 354A | Croker Hill | |
| 383 | Higher Minnend. | Evans et al. 42-43 (1968) |
| 410 | Old Hag | |
| 446 | Wiggenstall | |
| 461 | Easing | Morris 118 (Thesis 1966) |

CORRELATION WITHIN THE AREA STUDIED.

Morris (Thesis 1966) suggested that there are two horizons with Ct. edalensis in the succession separated by "some 400 feet of sandstones and shales." This was due partly to an error in mapping and partly due to his identification of the E. grassingtonense band at Loc. 001 (Morris 98) as that of C. subplicatum. The true identity of the fauna of Loc. 001 has been discussed (page) and it only remains to equate the Ct. edalensis faunas of Loc. 002 Hurdlow (Morris 119) and Loc. 461 Easing (Morris 118) with one another. The present mapping of the Hurdlow area clearly shows the Ct. edalensis horizon to lie within shales above the C unit of protoquartzites which overlies the E. erinense horizon (Figs. 1.1c-e). In the section at Easing the E. erinense horizon (Loc. 463) has been found below the C unit of protoquartzites and the Ct. edalensis horizons occur in the shale-mudstone overlying these protoquartzites. The presence of the L. longirostris horizon at the top of the protoquartzite unit and below the Ct. edalensis bands makes it certain that these two Ct. edalensis localities are at the same level and not separated by 400 feet of strata.

Holdsworth (Thesis 1963) considered that two horizons were present in the Upper Dove. The lower with C. cf. subplicatum, Ct. cf. bisati (rare) and Ct. sp. (Holdsworth Loc. 266), and the higher one with Ct. edalensis only

(Holdsworth Loc. 245). He tentatively suggested that these horizons are about 20 feet apart. The present author was unable to verify the fauna at Holdsworth's Loc. 266 but has examined the material obtained by Holdsworth and agrees that Cravenoceratoides is certainly present, but Cravenoceras is far more abundant. This fauna is six feet above the C. subplicatum solo fauna in the Upper Dove and has not been identified with certainty at any other exposures in the area. The only possible equivalent is a sparse fauna of Cravenoceras sp. obtained from a bullion 80cm. above the C. cf. C. subplicatum limestone at Loc. 214 Blake. The fauna at Holdsworth Loc. 245 (my Loc. 303) is clearly the normal Ct. edalensis fauna.

CORRELATION WITH OTHER AREAS.

The horizon of Ct. edalensis - Ct. bisati is a very widespread one and forms a valuable marker horizon from Ireland through Britain to the Continent.

On Slieve Anierin this horizon is the lowest one found by Yates (1962) above a grit which is about 200 feet thick and contains two coal seams. Below the grit there is a 100 feet of shale-mudstone with 'clay-ironstone bands' before the E. erinense horizon is reached. Thus the shales above the grit represent a re-extension of true marine facies to the area after the deltaic grit episode.

Hudson (1944) records the C. subplicatum fauna already mentioned from below a horizon with Ct. bisati in the sections of Hampsthwaite House, Coppice Beck and of the Simonseat Anticline. In these sections Ct. bisati is 10, 8 and 4 feet respectively above C. subplicatum.

In the Alport bore (Hudson and Cotton 1943) Ct. edalensis is recorded 7 feet above the fauna of Cravenoceras sp. and E. bisulcatum which is thought to lie close above the Eumorphoceras sp. nov. II horizon (see page).

In Edale it appears that there are several horizons present in this portion of E2b and that Ct. edalensis is not present in all of them. Forms from Loc. 180 of Hudson and Cotton (1945) (Loc. H.C. 180) are not Ct. edalensis, but those from Loc. H.C. 144 near Edale church undoubtedly are of this species. A further locality yielding true Ct. edalensis has been found by the author near Loc. H.C. 112 in a small tributary about 30 yards from its junction with the main stream at Loc. H.C. 112. At both these localities the Ct. edalensis horizon can be seen to be above the prominent pair of clay beds B5 and B6 and direct correlation with north Staffordshire is possible. The Ct. aff. edalensis faunas of Hudson and Cotton (1945) from Loc. H.C. 180 and H.C. 162 are considered by the author to lie between the C. holmesi Bisat and Ct. edalensis horizons and thus there is not considered to be a C. subplicatum horizon

exposed below the Ct. edalensis horizon in Edale.

Moseley (1954) records Ct. lirifer (= Ct. edalensis) and rare Ct. bisati in several bands within shale for 12 feet above the surface of the Roeburndale Grit in the Lancaster Fells. The lowest horizon may rest directly on the grit surface. Coals occur within the grit giving the succession a similarity to that on Slieve Anierin.

On the Continent Schmidt (1934) records Ct. edalensis from Arnsberg in Westphalia and Demanet (1941) records it from the Namur area in Belgium.

Explanation Fig. 1.3q

Fig. 1.3q

1. Cravenoceratoides edalensis (Bisat) x5, Spec. 354AA, Loc. 354A, Croker Hill.
2. Cravenoceratoides bisati Hudson. x5, Spec. 060A, Loc. 060, Dalehouse Wood.
Solid specimen showing irregular bifurcation of the lirae.

Both specimens are from the E2b1 Ct. edalensis faunal horizon.

Fig 1.3q



1
2



CRAVENOCERAS cf. C. SUBPLICATUM - SECOND HORIZON.

This horizon consists of isolated bullions of limestone in shale. It is known from three localities in the area but only one yields bullions with C. cf. C. subplicatum. At Loc. 002 Hurdlow the bullions occur 23cm. above the pyritic clay bed B7 and 540cm. above Ct. edalensis. They have yielded the following fauna:-

Dimorphoceras sp. abundant

Cravenoceras cf. C. subplicatum rare

P. corrugata (quadrate form) common

Fish scales

Orthocone Nautiloid

At the other two localities, Loc. 114 Pyeclough and Loc. 202 Blake, the position of the fauna can be found relative to pyritic clay bed B7, but no Cravenoceras have been seen, possibly due to the rusty nature of the shales which would obscure the delicate ornament of a Cravenoceras.

The presence of a C. cf. C. subplicatum fauna below that of Ct. edalensis has already been demonstrated, and correlated with similar horizons elsewhere. Yates (1962) records the maximum development of C. subplicatum above Ct. edalensis and this record of a horizon with C. cf. C. subplicatum above Ct. edalensis correlates with Yates' C. subplicatum band in a similar position above Ct. edalensis.

A horizon of bullions with ?Anthracoceras and Cravenoceras was recorded by Stephens et al. (1942) six feet above a level with Ct. aff. edalensis (Ct. bisati?) at March Ghyll reservoir on the southern flank of the Beamsley anticline. This horizon may possibly correlate with this second appearance of C. cf. C. subplicatum.

EUMORPHOCERAS aff. LEITRIMENSE AND CRAVENOCERATOIDES
aff. NITIDUS HORIZON.

This horizon was found in the Pyecclough Brook at Loc. 113 which is about 15 yards up a small sidestream running into the main brook about 30 yards downstream of the bridge. The specimens occur in very weathered shale and are extremely difficult to preserve. The specimens of Ct. aff. nitidus were found in a few centimetres of dark grey shale and the Eumorphoceras specimens were found in about 30cm. of green weathering shale immediately below this horizon.

The stratigraphic position of this fauna is above the Ct. edalensis fauna and separated from it by about 13 metres of shale and shale-mudstone. The fauna is also above the topmost pyritic clay bed B7 (see section 1.4) the distance being about 7 metres. No faunas with distinctive goniatites have been found between Ct. edalensis and this faunal horizon at this locality, but from the Hurdlow 002 section C. subplicatum is recorded just above the pyritic clay bed B7 (page). Upwards in the succession the next well marked fauna seen in the Pyecclough Brook is that of the E2b3 cherts which contain Ct. nititoides (Bisat) in this area (Holdsworth 1963). The relations with the C. holmesi bands which occur below the cherts (Holdsworth 1963)

cannot be determined with certainty since C. holmesi has not been found in this section, but C. holmesi is suspected by the author to be above the fauna under discussion.

FAUNAL DESCRIPTION.

Family	GONIATITIDAE	de Haan
Subfamily	GIRTYOCERATINAE	Wedekind
Genus	EUMORPHOCERAS	Girty 1909

Eumorphoceras leitrimense Yates

Eumorphoceras bisulcatum leitrimense Yates 1962 p.385-6

Pl.53 Figs.4,5. Pl.54 Figs.5,6. Pl.55 Fig.5

Eumorphoceras aff. leitrimense Yates Fig.1.3r,1.3s

Material. 7 specimens crushed in shale.

Locality. Loc. 113 Pyeclogh Brook.

Description.

The specimens are all small, the greatest diameter at the spiral groove being 7mm. (Fig.1.3s), at this diameter there are 12-14 ribs per $\frac{1}{2}$ whorl. The ribs are rather short often not reaching the spiral groove and are strongest near the umbilicus. The ribs are radial for most of their course, but $\frac{2}{3}$ of the way to the spiral groove they turn sharply forward and die out before reaching the groove. At a spiral groove diameter of 7.5mm. the umbilicus is 2.8mm. wide. Spec.113H (Fig.1.3r2)

is small, 3.5mm. diameter, but shows two constrictions with about 5 ribs between. Several fragments were collected showing only a spiral groove. This is probably an old age form.

Discussion.

As is mentioned by Yates (1962 p.386) in her discussion of this species the only form described which in any way resembles E. leitrimense is E. bisulcatum varicata Schmidt, but this form has a wide umbilicus in Schmidt's figure and there is no sign of the ribs dying out at 12mm. diameter. Also there are more ribs between constrictions in varicata (7 reducing to 4) than in leitrimense (3 reducing to 2).

The present material is similar to Yates' subspecies in having a fairly small umbilicus, but in the only example where constrictions are clearly seen there are 5 ribs between. This specimen is smaller than those illustrated by Yates and so this feature cannot afford evidence that the specimens described here are not E. leitrimense since in both this species and in varicata the number of ribs between constrictions reduces with increase in the size of the shell. The ribs are similar to those illustrated by Yates and about the same number occur, e.g. $12\frac{1}{2}$ whorl on Yates 1962, Pl.53 Fig.4. The specimens described here are referred to E. aff. leitrimense.

Subfamily	HOMOCERATINA	Spath
Genus	CRAVENOCERATOIDES	Hudson 1941

Cravenoceratoides nitidus (Phillips)

Goniatites nitidus Phillips 1836. p.235-236 Pl.XX Figs.10-12

Homoceras nitidum (Phillips) Bisat 1924 p.106

Cravenoceras nitidum (Phillips) Bisat 1932 p.34-35 Pl.II

Fig.3

Cravenoceratoides nitidus (Phillips) Hudson 1946 p.376,383

Pl.XXI Fig.11 Pl.XXIa Figs.1a-c

Cravenoceratoides nitidus (Phillips) Yates 1962 p.390-1

Pl.52 Fig.6

Cravenoceratoides aff. nitidus Phillips Fig.1.3t

Material. About 20 crushed specimens in very friable
black shale, mainly external moulds.

Locality. Loc. 113 Pyecrough Brook.

Description.

The largest specimen has a crushed diameter of 15mm.
and an umbilicus 4.5mm. wide. There appears to be a
fairly strong umbilical shoulder present and the edge
of the umbilicus is taken at the shoulder in the above
measurement (Fig.1.3t1).

Well developed lirae are present that are canted
and arise in the umbilicus and pass radially over the
umbilicus. On the flank the lirae may bend slightly
forward but appear to be almost straight over the venter.

On the flanks from 1 to 3mm. from the umbilical edge a second set of lirae is intercalated between the existing lines (Fig.1.3t2). The lirae do not bifurcate as in Ct. edalensis since the lirae are not actually joined. The intercalated lira often rises near another lira and diverges from it, gaining strength, for a short distance before they run parallel and are of equal strength.

Discussion.

These specimens are easily distinguished from the earlier Ct. edalensis and C. bisati since they possess canted and not symmetrical lirae, a fact mentioned by Hudson (1946 p.385).

Ct. nititoides is distinct from the present material in that it has a much smaller umbilicus and the lirae have a much stronger forward curve which is at its strongest near the umbilical margin. The intercalation of a second series of lirae is however similar to that seen in the present material.

Nuculoceras (Ct.) stellarum (Bisat) is distinct in possessing a concentric sculpturing of the lirae a feature noticed by Holdsworth (1965 p.228) which places it in the genus Nuculoceras.

Ct. nitidus was first described by Phillips (1836) but by present day standards the description is not good enough for comparison with the specimens described here.

Bisat (1932) redescribes material from Dinckley on the River Ribble which he says agrees closely with Phillips type material in the British Museum. The material from Dinckley was however from a loose bullion, and so its exact stratigraphic position is unknown.

The umbilical size of the present material agrees well with that mentioned by Bisat - 4mm. at 12mm. diameter, but he describes the lirae as 'dichotomising' and although this is nearly the case, the secondary lirae on the present material do not actually branch from a lira but are intercalated between the other lirae. Examination of Bisat's material shows that the ribs are in fact identical to those in the present material, which appears to be conspecific with Bisat's. Yates (1962) also mentions that the lirae bifurcate, but in her figure some if not all the lirae appear to be intercalated. Ct. nitidus is certainly the nearest to the present material of all the described species of Cravenoceratoides.

OTHER LOCALITIES WITHIN THE AREA.

The only other possible measured section in which this horizon would be expected to occur is that of Hurdlow Loc. 002. At this locality the pyritic clay bed B7 occurs 520cm. above the top of the Ct. edalensis horizon, and 440cm. above B7 a thin shaley limestone yielded P. corrugata in abundance and two fragments of a

Cravenoceratoides with canted lirae. This fauna may be the equivalent of the Loc. 113 Pyecrough fauna since both mark the first introduction of Cravenoceratoides of nitidus-nititoides type and hence the base of E2b2.

CORRELATION WITH OTHER AREAS.

Yates (1962) records E. leitrimense at the base of the Ct. nitidus zone since she finds associated with it fragments of Cravenoceratoides with canted lirae as found in Ct. nitidus and Ct. nititoides. She records rare C. holmesi 'slightly lower' than this horizon but says it is more abundant above. This horizon of Yates would seem to correspond with the one in the present area. Yates (1962) correlates this horizon with the Colsterdale Limestone of the Greenhow Mining Area which contains Ct. nitidus and rare 'E. bisulcatum' which is characterised by short ribs. This is in turn correlated with the Birk Gill Limestone of the Simonseat Anticline (Hudson 1939) and the nitidus limestone of the Lancaster Fells described by Moseley (1954).

A fauna recorded by Earp et al. (1961 p.124) in the Clitheroe Memoir from a stream (966438) near Stubbing includes a Cravenoceratoides described as being intermediate between Ct. nititoides and Ct. fragilis. Specimens examined from this locality by the author at the Institute of Geological Sciences, Leeds, are identical with the Ct. nitidus described above from the present area. At Stubbing

the Cravenoceratoides is associated with a Eumorphoceras described in the Memoir (p.124) as E. bisulcatum. Examination of these specimens shows them to be E. leitrimense. This horizon is probably the same as the one found in the present area.

Ct. nitidus has been recorded by Hudson and Cotton from both Alport (1943, p.162) and Edale (1945, p.9). The faunal band referred to in each case is the E2b3 faunal band which also contains Ct. nititoides, E. rostratum Yates, brachiopods and trilobites. This faunal band is characterised by chert development both in north Staffordshire and in Edale and is not the same horizon as the Ct. aff. nitidus horizon described here. In fact the E2b3 horizon is present in Pyeclough Brook at Loc. 116 and lies above the horizon with Ct. aff. nitidus and E. aff. leitrimense.

Explanation Figs.1.3 r-t.

Fig.1.3r.

1. Eumorphoceras aff. leitrimense Yates. x15, Spec.113A.
2. Eumorphoceras aff. leitrimense Yates. x15, Spec.113H,
small specimen with constrictions.

Fig.1.3s.

Eumorphoceras aff. leitrimense Yates. x10, Spec.113B.

Fig.1.3t.

1. Cravenoceratoides aff. nitidus (Phillips). x7.5, Spec.113L.
2. Cravenoceratoides aff. nitidus (Phillips). x10, Spec.113K.

All specimens shown in Figs.1.3r,s and t are from
Loc.113, Pyecclough.

Fig 1.3r



1

2

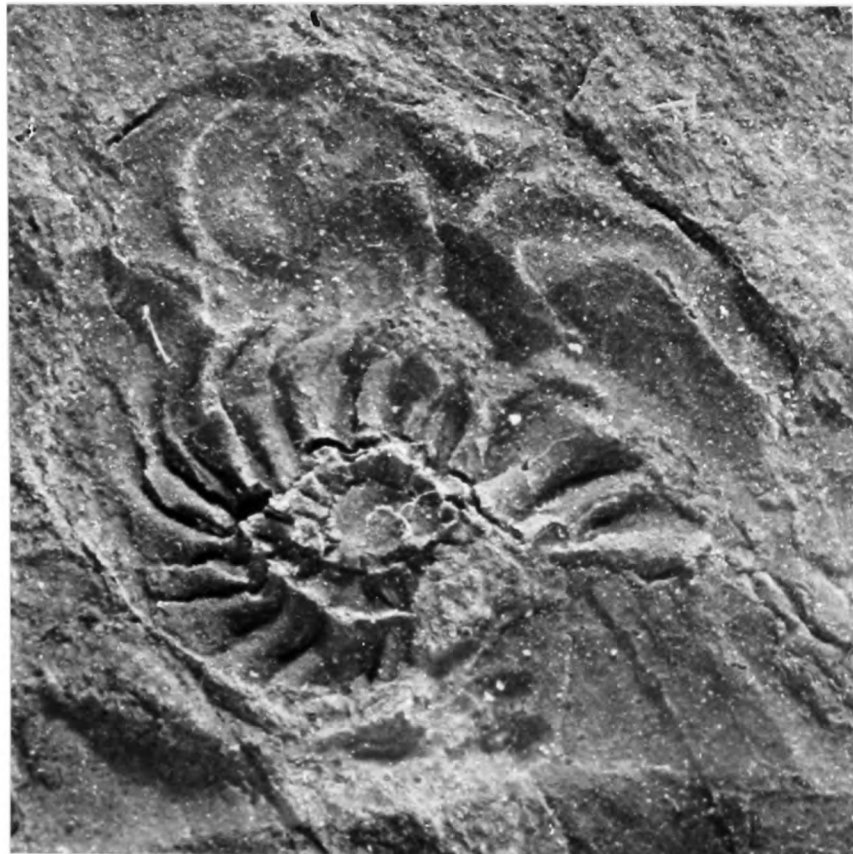


Fig 13s

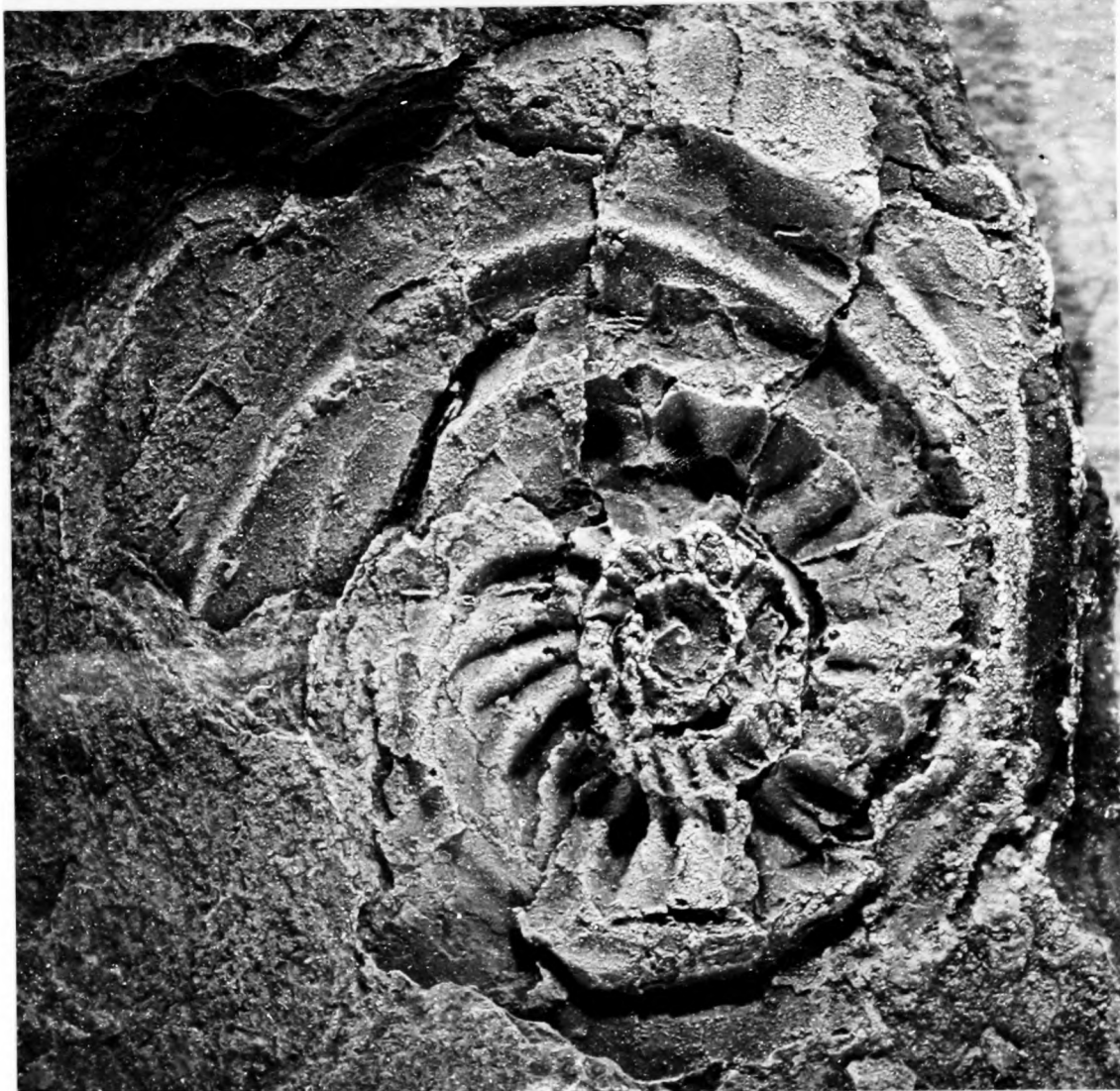
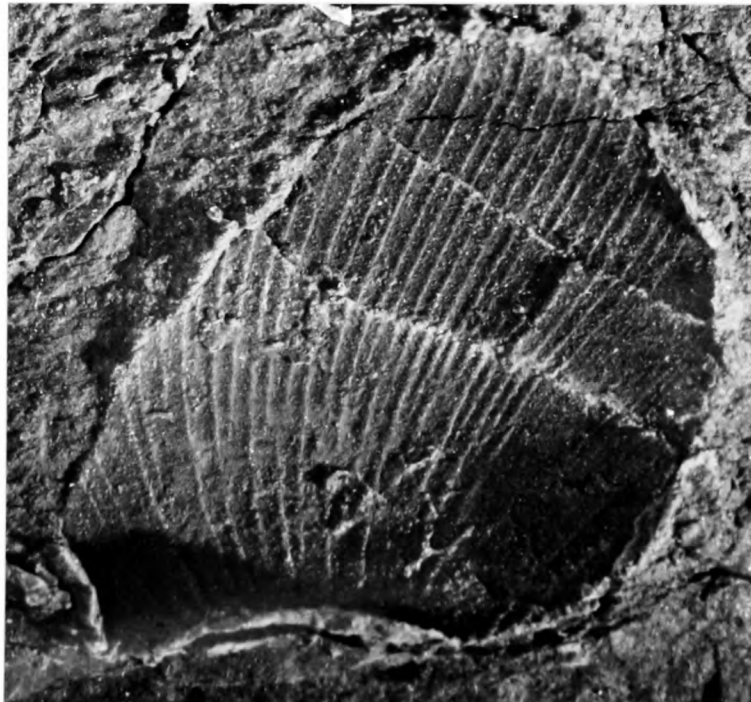


Fig 13t



1

2



PALAEOLOGICAL CONCLUSION.

The major marine bands with C. cf. C. malhamense E1c, E. grassingtonense E2a1, E. erinense and E. ferrimontanum E2a2 and Ct. edalensis E2b1 are present over the whole of the basin area in north Staffordshire, and north of the Derbyshire massif in Edale. The fauna of each band is distinctive and does not appear to vary significantly within the area studied. Above Unit C of the proto-quartzites several faunal bands in addition to the Ct. edalensis band are present and are useful for detailed correlation.

The sequence of marine bands and faunas recognised agrees well with that described by Yates (1962) from Slieve Anierin in Eire, and the marine bands provide accurate correlations of the succession over wide areas. In the north Staffordshire area the marine bands were found to maintain constant positions relative to the coarser grained lithofacies and to the K-bentonite clay beds described in the next section of this thesis.

SECTION 1.4.

THE K-BENTONITE CLAY BEDS.INTRODUCTION -

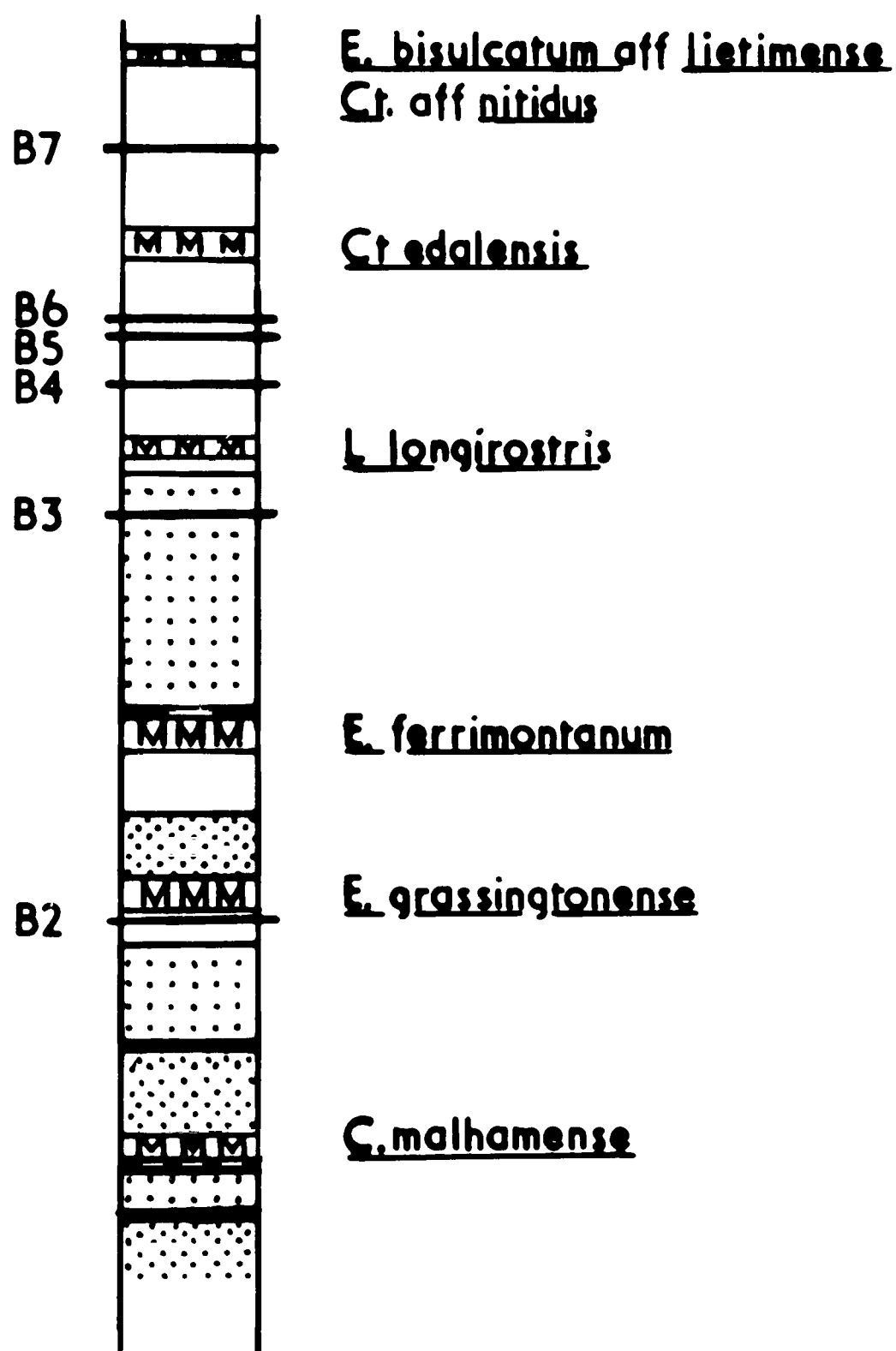
During the study certain thin distinctive clay beds were noticed at several horizons in the succession. These beds were found to maintain their positions relative to the goniatite horizons in all exposures where they occurred. The relative positions of the clay beds and the faunal bands are shown in Fig. 1.4a.

The tight faunal control supplied by the goniatites in the Upper Pendleian and Lower Arnsbergian of the Namurian, and the proximity of the thin clay beds to the goniatite horizons enables the absence or presence of a clay bed to be determined with reasonable certainty. Accurate correlation of beds from one exposure to another can be made in the area.

The clay beds are all less than 5 cm. thick, but usually show up well in an outcrop since they are pale in colour compared with the dark shale-mudstone in which they occur. The actual colour of the beds can be yellow or orange due to oxidised iron, or pale green, blue or white. Thus there is generally no difficulty in the identification of a bed, the exception being when a bed has been secondarily calcified as has occurred at Loc. 202

Fig 1-4a

RELATION OF K BENTONITES TO FAUNAL
HORIZONS IN N STAFFS BASIN



(Blake Brook). Then it takes on a grey colour and is more difficult to find, and would almost certainly be missed if its presence was not suspected. Nearby at Loc. 201 this bed is not calcified and has a normal yellow appearance.

THE INDIVIDUAL CLAY BEDS.

The clay beds have been designated B1-B7 and will be described individually. B1 only occurs in Edale, Derbyshire, and is described later.

Bed B2.

This bed has been found in every outcrop of the horizon where it would be expected to occur. It is generally 0.5 cm. thick and yellow to orange in colour due to the oxidation of the iron pyrites by which it is often partially replaced. The fact that this is a single thin bed of great lateral extent is shown by the table of measured distances of the bed above the B unit of protoquartzites, and below the base of the E. grassingtonense faunal horizon. The consistency of the measurements is striking, the bed remaining roughly in the middle of the interval from the topmost sandstone to the base of the faunal band.

Locality.		cm. above top sst of B unit.	cm. below <u>E</u> <u>grassingtonense</u> band.
Hurdlow	008	120	150
Hurdlow	001	145	155
Hurdlow	009	110	-
Elkstones	505	120	90
Pyecrough	120	135	125
Blake	209	90	90
Winkhill	581	150	90
Sunnydale	252	120	130
Croker	357A	150	155
Sprink	050	110	160

Bed B3..

This horizon has only been recognised at Pyecrough (30 yds. upstream from 101) in the present area. It is 0.5 cm. thick and is not pyritic, consisting only of a pale bluish clay often stained yellow or orange at the margins of the bed. B3 occurs within the C unit of protoquartzites 10 metres below the L. longirostris horizon. At Pyecrough it occurs in shale-mudstone between protoquartzites and siderites. The bed could easily have been eroded from localities nearer the source of the protoquartzites accounting for the fact that it has only been seen in this one exposure in the area.

Bed B4.

This bed has been found at three localities in the area, Wigenstall 446, Pyecrough 101 and Old Hag 411. It is 0.5 cm. thick at all localities. Pyrite occurs as patches in the bed but not to the same extent as in B.2. The bed occurs in the interval between B5 and the L. longirostris horizon. At Pyecrough it is also above the Eumorphoceras sp.nov.II horizon.

Beds B5 and B6.

These two beds are the most easily noticed of the claybeds. They are generally about 20 cm. apart and are set in unfossiliferous shale-mudstone. B6 is the thickest and apparently the most widespread of the two, being found at all suitable localities in the area, although it is reduced to a green rusty parting at Blake Locs. 201 and 202. B5 is not so extensive being absent at Croker 354 and Higher Minnend 380. At the five other localities exposing this horizon both B5 and B6 are present.

These beds occur beneath the Ct. edalensis horizon, and above L. longirostris and Eumorphoceras sp. nov.II. The beds vary in colour from orange-yellow to pale blue and do not contain any pyrite visible in the hand specimen. Fig. 1.4b shows the relations of B4, B5 and B6 to the adjacent faunal bands.

Fig 1.4b

LOCALITY AREA AND NUMBER	Cms.ABOVE L.longirostris BAND OF B4, B5 OR B6	B 4	Sh. Mdst. between B4 & B5	B 5	Sh. Mdst. between B5 & B6	B 6	Cms.BELOW Ct. eddensis BAND OF B6
WIGGENSTALL 446-7	560	0.5	320	1.5	29	2.0	820
PYECLOUGH 101	430	0.5	120	1.4	26	1.5	750
OLD HAG 411	—	0.5	120	1.0	24	1.5	—
BLAKE 201	540			1.0	23	P†	520
HURDLOW 002	270			0.8	21	1.2	400
EDALE 112	420			4.0	18	1.0	c 400
THORNCLIFF 421	c 200			0.5	17	1.0	c 300
MINNEND 382	290					0.5	—
CROKER 354	280					1.0	610

Bed B7.

This bed has been recognised at three localities in the area, Blake 202, Pyecrough 114 and Hurdlow 002A. At all three localities it is in shale that contains a fauna which includes Posidonia corrugata, Posidoniella sp., A.-D. sp., and at Hurdlow C. cf. C. subplicatum orthocone nautiloids and fish scales in addition to the usual fauna.

This horizon occurs above Ct. edalensis and in the Pyecrough section is below the horizon with Ct. aff. nitidus and E. aff. lietrimense. The position of the bed relative to B6 and the top of the Ct. edalensis band is shown below.

Locality	thickness cm.	distance above <u>Ct. edalensis.</u>	distance above B6.
Blake 202	1.5	675	1325
Pyecrough 114	1.3	600	1350
Hurdlow 002A	1.0	520	1070

These figures are thought to indicate real differences in the thickness of the shale at different localities. The continuous exposure found in the cliff at Blake 202 ensures that there is no other clay band within this interval. In appearance the bed is usually yellow or orange and rich in pyrite as in B2.

THE COMPOSITION OF THE CLAY BEDS.

The clay mineralogy of the beds was investigated

using samples of B2, B5, B6 and B7 obtained from several different localities so that comparisons could be made between individual beds and also between different outcrops of the same bed. Samples of the shale and the shale mudstone in which the clay beds occur were also taken for comparison. The specimens were examined by means of a Siemens X-Ray diffractometer using Cu K α radiation.

SAMPLE PREPARATION.

The collected specimens were first dried and then ground down and passed through a 240 mesh sieve. From 0.5 to 1.0 gms. of the ground material was placed in a 50 ml. beaker with 10-20 cc. of distilled water and a few drops of 1% Calgon solution to prevent flocculation of the clay particles. Dispersion was aided by suspending the beaker with clay and water in an ultrasonic vibrator for about one minute. The beaker was then allowed to stand for about fifteen seconds before some of the suspended material was transferred to a glass slide by means of a pipette, and allowed to dry slowly by evaporation of the water at room temperature. This method of sedimentation of the clay produced a reasonably flat and thin coating on the glass slide with fair orientation of the clay minerals. Other methods of preparation of the samples were tried but were found to be less successful. These methods included sedimentation of the clay from industrial

spirit, and stirring the clay material on a slide with a few drops of water to produce a slurry which was then allowed to dry. Both methods gave a poorer orientation of the clay minerals and a more uneven surface to the specimen resulting in poorer peaks and higher background on the diffractometer trace. Three slides were prepared of each specimen from the same clay/water mixture to obtain relative uniformity. One slide was examined without further treatment, the second after glycolation for one hour at 60°C, and the third after heating at 550°C for two hours.

DIFFRACTOMETER SETTINGS.

Each specimen was scanned from 3-30° 2θ with controls set as follows:-

Generator	35kv	20mA
Slits	.3mm	.1mm
Baseline	11.0	Channel width 23.0
Detector	1985v	
Statistical error	4%	
Imp/Min	1x10 ⁴	
Scanning speed	1° 2θ/Min	
Paper speed	1 cm/Min	

The controls were kept at the same setting for all the specimens. Some specimens from which only confirmatory results were required were run from 3-15° 2θ to cover the

main clay mineral peaks.

DIFFRACTOMETER RESULTS.

The peak positions for the clay bed specimens and shale-mudstone specimens are shown in Figs. 1.4d and 1.4e. The positions of the shale-mudstone samples relative to the clay bands are shown in Fig. 1.4c, and typical diffractometer traces are shown in Figs. 1.4f,g,h (clay bed), and Fig. 1.4i shale-mudstone.

The results show that there is a distinct difference in the clay mineralogy of the clay beds and of the shale-mudstones in which they occur. There is little variation between the different clay beds, certainly not enough to identify individual clay beds without stratigraphical evidence.

DESCRIPTIONS OF TRACES FROM INDIVIDUAL CLAY BEDS.

Bed B2. Specimens 13 and 17.

The traces are rather poor with the main clay peak broad and not very sharp. The kaolinite peaks are sharp. The background level is high possibly due to the presence of pyrite and iron hydroxides. Specimen 17 contains quartz and is the only sample of a clay bed found to contain quartz. There is no quartz in specimen 13. Trace of specimen 13 is illustrated, Fig 1.4f.

Bed B3. Not investigated.

Bed B4. Not investigated.

Fig 1.4c

RELATIVE POSITIONS of CLAY BED and SHALE-
MUDSTONE SAMPLES

PYECLOUGH HURDLOW THORNCLIFF CROKER

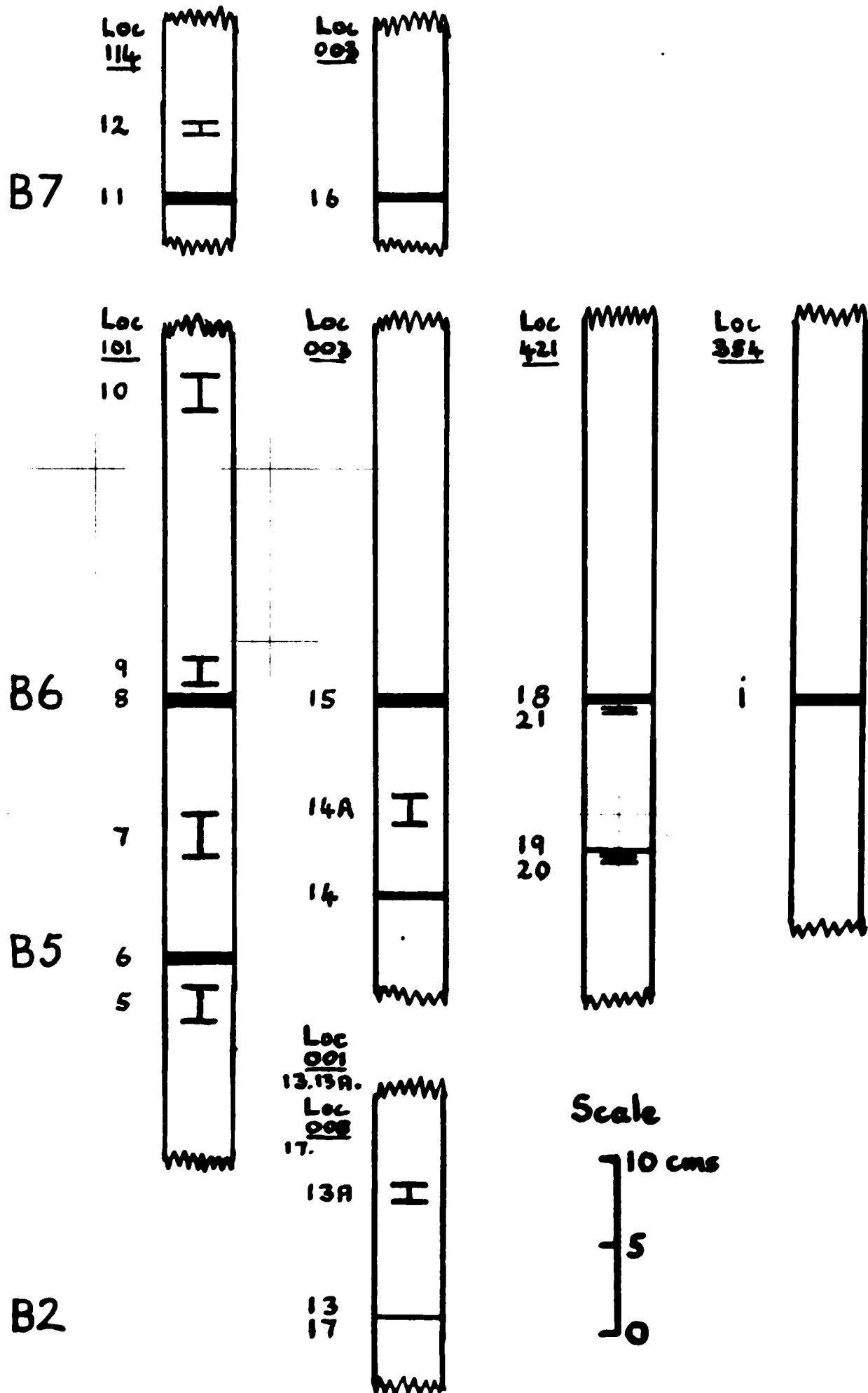


Fig 1.4d

K-Bentonite Clay Beds

Peak positions from diffractometer traces (in Å).

Spec & Bed Nos		MM 001, 001	K 001	MM	MM	H 002	MM
13	O	11.2	7.13	5.11	4.47	3.58	3.29
	G	11.5	7.20		4.47	3.59	3.35
B2	H	10.04			4.49	(3.59)	3.38 3.32
17	O	12.2	7.18	5.02	4.46	3.60	masked by
	G	(14.7-12.6) (10.9-9.7)	7.21	5.00		3.59	quartz
B2	H	10.03		4.99	4.51		reflection
	O	11.9	7.18	5.06	4.48	3.58	3.34
	G	12.3 10.4-9.5	7.16	5.03	4.49	3.58	3.36
B5	H	9.93		4.90	4.49	(3.58)	3.27
14	O	12.3	7.27	5.07	4.47	3.59	3.35
	G	12.7 10.3-9.1	7.21		4.49	3.59	3.38
B5	H	10.10		4.99	4.51	(3.53)	3.30
19	O	11.8	7.17	5.06	4.49	3.58	3.34-3.14
	G	12.9 9.8-9.3	7.21	5.31	4.50	3.58	3.35
B5	H	10.02		4.97	4.49	(3.55)	3.35
	O	12.0	7.18	5.12	4.56	3.59	3.20
	G	12.9 10.6-9.4	7.21	5.24	4.48	3.59	3.35 3.19
B6	H	9.93		4.93	4.49		3.27
15	O	11.9	7.16	5.01	4.47	3.59	3.35-3.20
	G	12.2 9.9-9.2	7.18		4.51	3.59	3.33
B6	H	9.96		4.96	4.49		3.27
18	O	12.0	7.20	5.07	4.49	3.57	
	G	12.6 10.3-9.3	7.20	5.29	4.47	3.58	3.35 3.13
B6	H	10.02		4.96	4.49	(3.53)	
1	O	11.9	7.17	5.04		3.58	3.38-3.18
	G	12.7 10.3-9.4	7.15	5.23		3.59	3.33
B6?	H	9.97		4.93	4.47		3.29
11	O	11.9	7.19		4.48	3.58	3.19
	G	13.6 12.8-11.8	7.15		4.46	3.59	3.35 3.20
B7	H	10.16			4.51		3.26 3.21
16	O	12.0	7.19		4.48	3.58	3.33 3.19
	G	12.6 (10.8)	7.15		4.46	3.58	3.33 3.19
B7	H	10.04		4.89	4.50		3.33-3.18

Brackets indicate weak reflections.

Hyphens indicate broad reflections between indicated values.

O = Ordinary trace. G = Glycolated. H = Heated.

MM = mica Montmorillonite reflections. K = Kaolinite reflections.

Fig 1-4e

Shale-Mudstones

Peak positions from diffractometer traces (in Å).

Spec No.		CH 001	ML 001	M 001	K 001	L 002	M 002	K 002	CH 004
5	O G H	14.7 13.9	11.0 (11.6) (11.8)	10.16 10.01 10.13	7.17 7.17	5.04	4.51	3.586 3.578	
7	O G H	13.8 14.0 13.9	11.2 (12.1)	10.6 9.98 10.07	7.18 7.17	5.02 5.02	4.46 4.49 4.49	3.578 3.581	3.550 3.528
9	O G H	(17.3) 14.1 14.5 (16.0)	11.5 -----	10.1 -----9.8 10.04	7.19 7.21	NOT DETERMINED			
10	O G H	13.9 -13.8- (14.2)	11.8 11.4	10.10 10.02 10.07	7.16 7.21				
12	O G H	(14.1)	11.2	10.07 10.07	7.21 7.19				
13	O A G H	14.1	10.9	10.04 10.07 10.07	7.21 7.19				
14	O A G H	14.0 14.0 13.8	11.3 11.8	10.03 9.96 9.96	7.16 7.20	5.01 4.97 5.04	4.48 4.49 4.50	3.575 3.578	3.521 3.521
20	O G H	(14.5) 14.1	(12.4) -----	(9.96) 10.03 10.02	7.15 7.18	4.93		3.586 3.586	3.528
21	O G H	14.1 14.2 13.8	11.2 (11.9)	9.98 10.04 10.04	7.18 7.18	5.00 5.00 5.03		3.578 3.578	3.535

Brackets indicate weak reflections.

Hyphens indicate broad or reflections between values indicated.

M = Muscovite., ML = Mixed layer clay. CH = Chlorite. K = Kaolinite.

O = Ordinary. G = Glycolated.

H = Heated.

Fig I-4f

K Bentonite B2 Spec 13

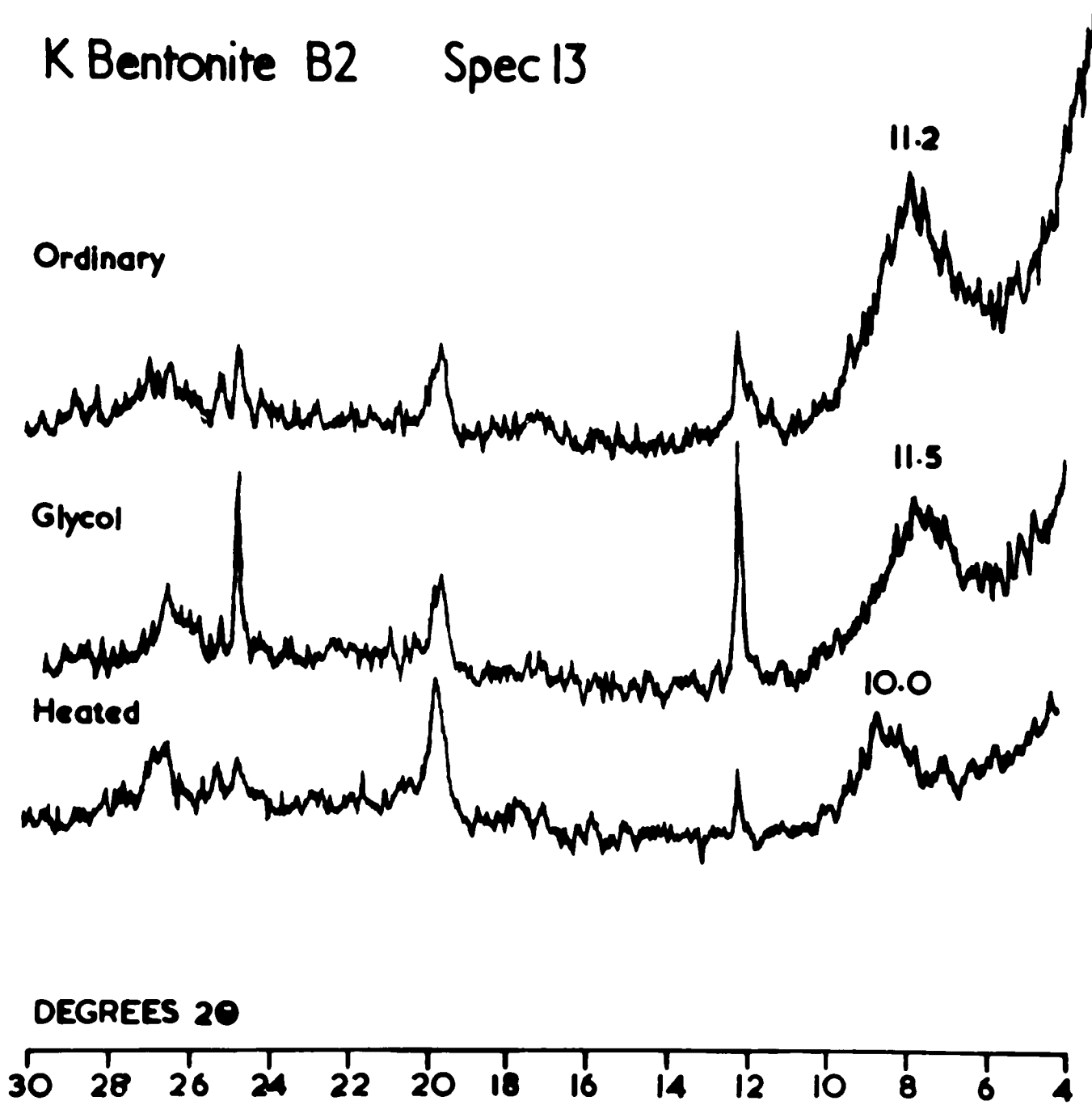


Fig I-4g

K Bentonite B6 Spec 18

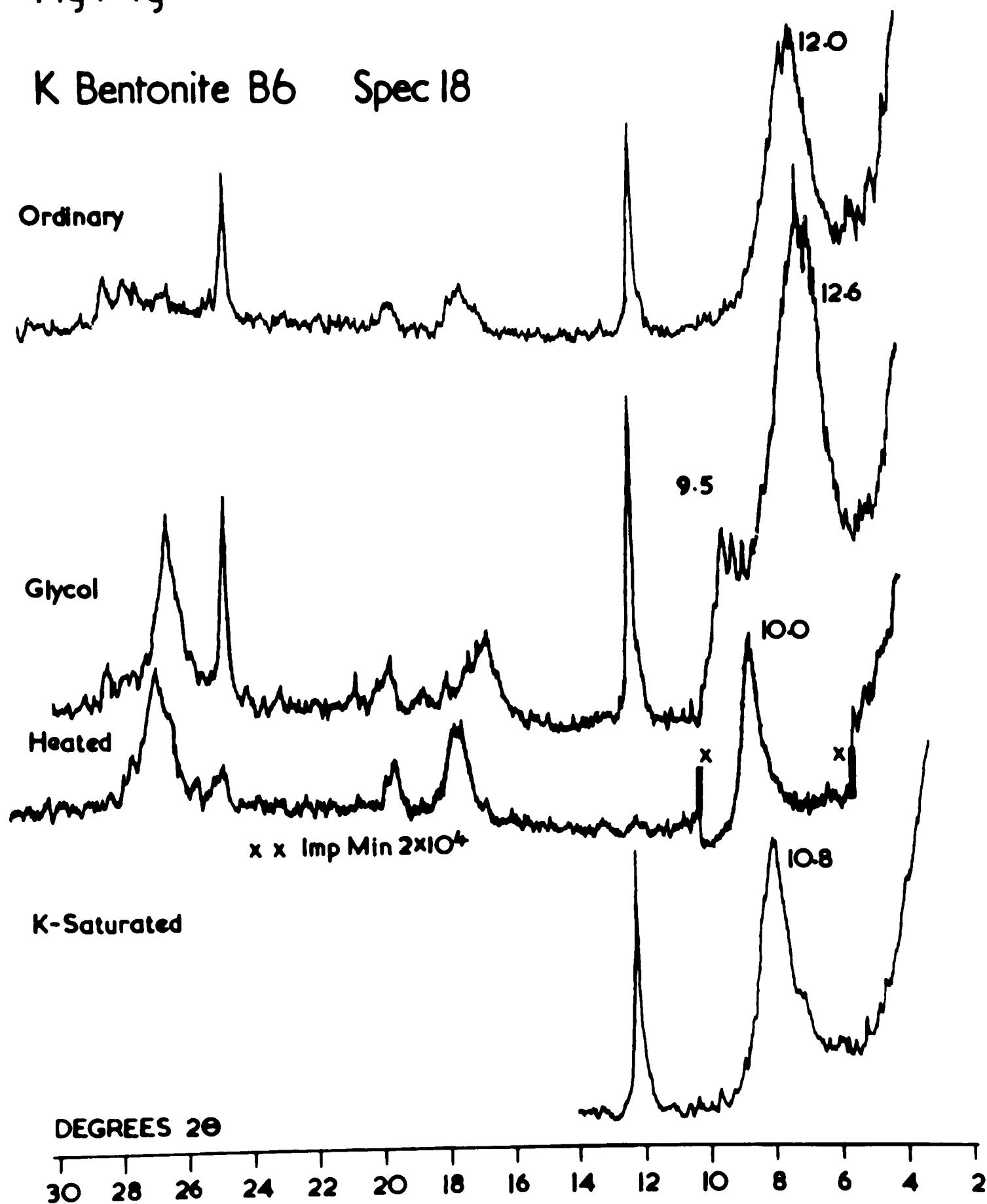


Fig 1-4h

K Bentonite B5 Spec 6

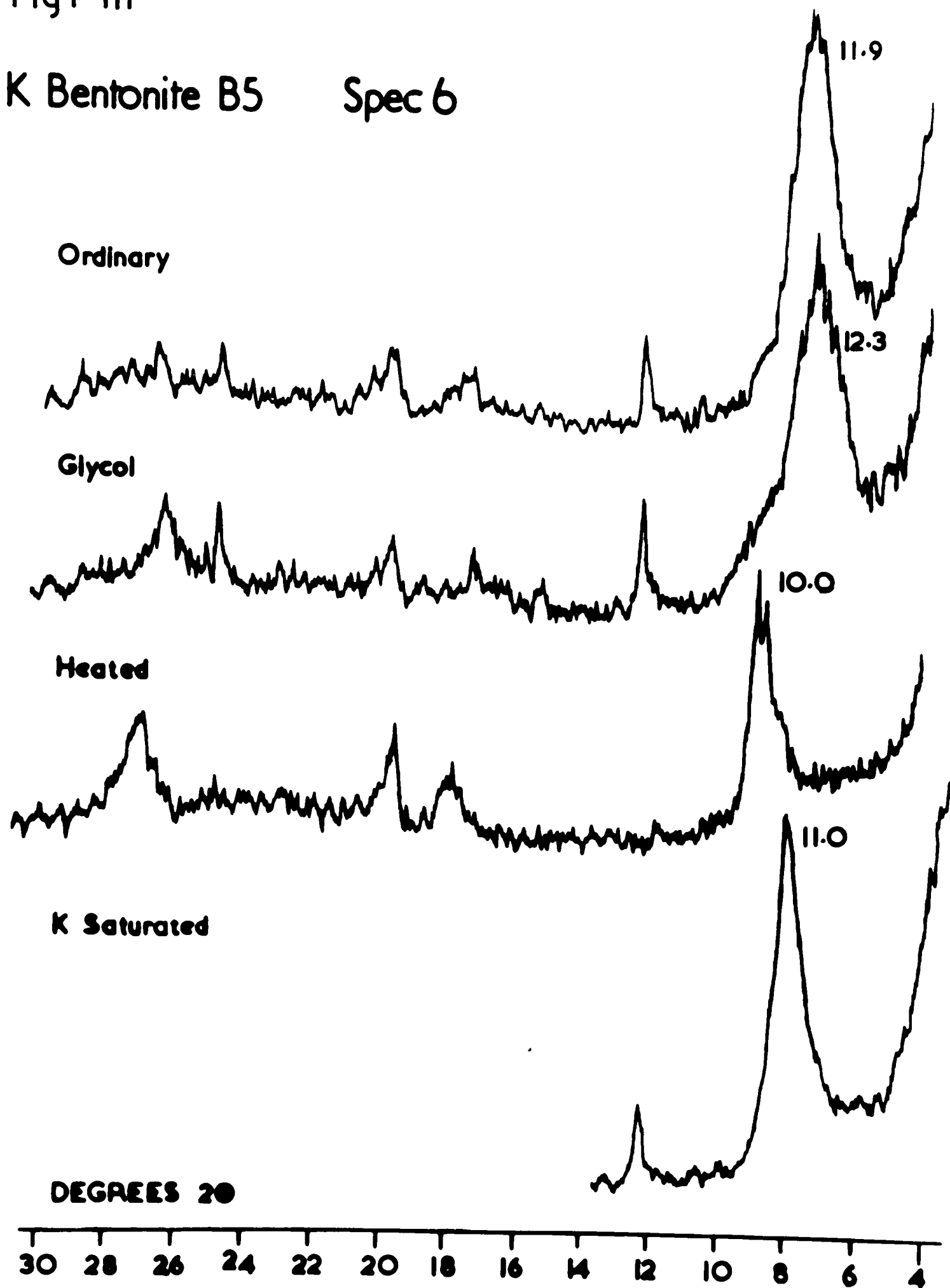
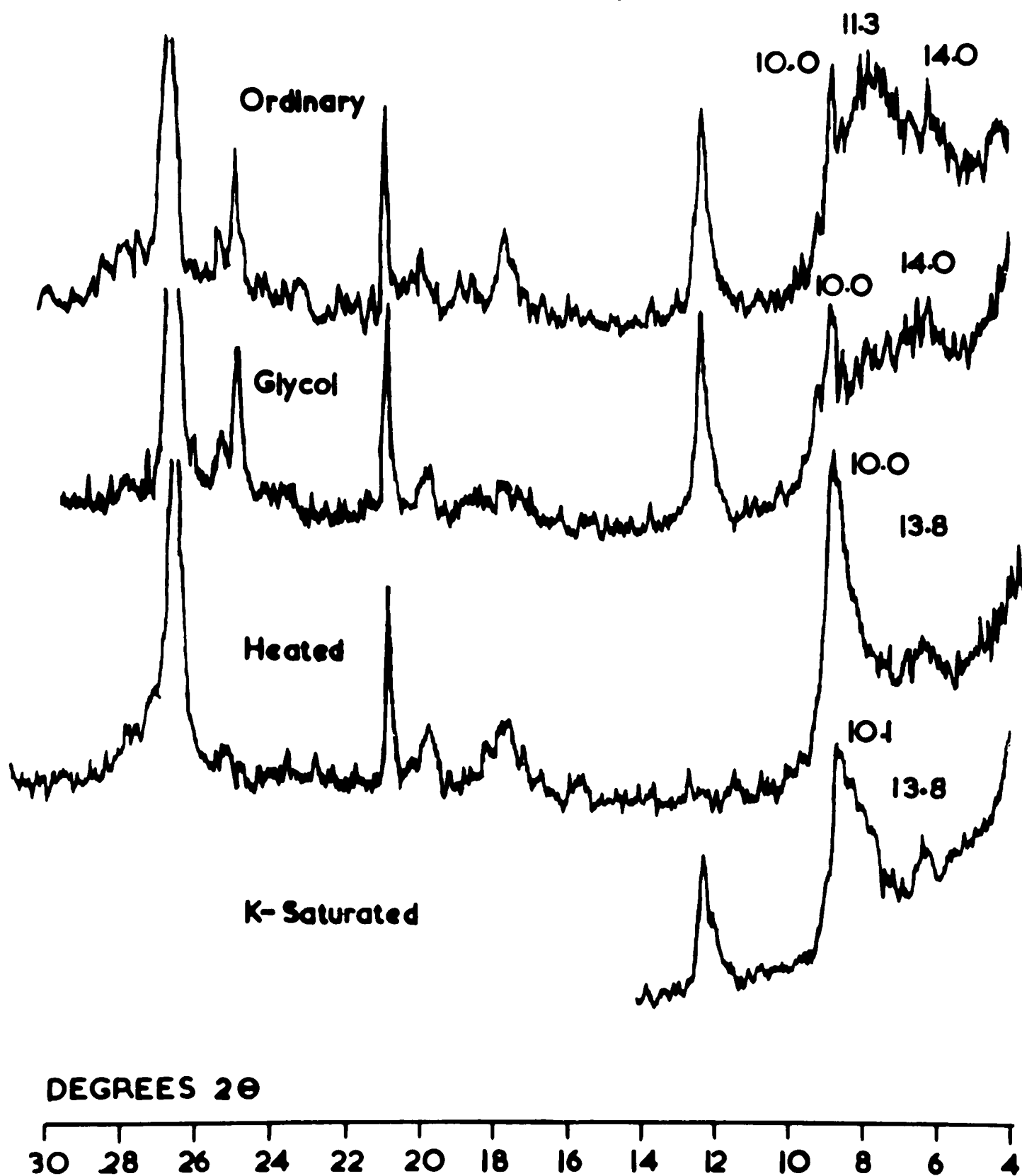


Fig 1-4i

Shale mudstone Spec 14a



Bed B5. Specimens 6, 14, 19.

All traces show strong mixed layer clay mineral peaks which vary in position from 11.35\AA to 12.26\AA . This variation may be due to various states of hydration in different samples or varying proportions of different layers caused by different states of weathering. The constancy in the position of the kaolinite peaks, which are always sharp, shows that the variation is real and not due to the apparatus. The two extreme values for the mixed layer peak were obtained from specimens done on different days and the variation in peak position may in part be due to changes in the relative humidity. A typical trace of this bed is illustrated, Fig. 1.4h.

Bed B6. Specimens 8, 15, 18, and ?1.

The traces produced from these specimens are very similar to those of B5, having a strong mixed-layer clay peak ranging from 11.87\AA to 12.04\AA , and also containing kaolinite. The only differences observed in the traces of B5 and B6 are in the relative amounts of kaolinite present, and the strength of the 001/002 mixed layer peak on glycolation. The following table shows the heights of the kaolinite peaks relative to the main mixed layer clay peak, and it can be seen that the height of the kaolinite peak in B5 is less than in B6 relative to the main mixed layer peak. The lack of more data makes the

distinction very tentative. The other minor difference is that the 001/002 peak of the mixed layer clay is stronger on the B6 specimens than on those from B5.

Bed.	Specimen.	Height of mixed layer peak above background.	Height of kaolinite 001 above background.
B5	6	12.8	2.8
B5	14	9.4	2.7
B5	191	13.0	2.8
B5	1911	11.2	4.3
B6	8	11.7	4.1
B6	15	12.6	6.7
B6	18	10.0	6.7
B6?	1	12.2	6.9

Specimen 1 (Croker 354) is either B5 or B6 from its position in the succession (see Fig. 1.4b). On the basis of kaolinite peak height and strength of the 001/002 glycolated peak it would appear to be B6. This correlation is strengthened by the observations that B4 is usually thicker than B5 in north Staffordshire and tends to maintain its thickness better. Thus the bed at Croker, and Higher Minnend is correlated with B6. The trace of specimen 18 is illustrated as typical of this bed, Fig 1.4g.

Bed B6. Specimens 11 and 16.

The specimens from this bed give similar traces to

those of B2 in having a broad, low mixed layer clay peak and also kaolinite peaks. The background level is very high c.20 relative to a maximum peak height of 47. This could be due to poor sedimentation and a lower proportion of clay due to the concentration of secondary pyrite in the bed, or to masking by iron hydroxides. The peak positions are however the same as in the other beds although some detail is probable obscured by the high background.

INTERPRETATION OF PEAKS AND CLAY MINERALS PRESENT.

1. THE CLAY BEDS.

It can be seen from Fig. 1.4d that the untreated clay bed samples all give a single broad peak in the range 11.2\AA - 12.25\AA with an average value of 11.95\AA . On glycolation there is usually a small expansion, nine of the eleven samples showing expansions averaging $.55\text{\AA}$. The two samples that only doubtfully show an expansion are 17 and 11, one from each of the pyritous bands. On heating for two hours at 550°C the clay lattice collapses to near 10\AA .

The mineral represented by this main 12\AA peak is thought to be a mixed layer clay mineral consisting of mica with some expandable layers of montmorillonite which are responsible for the expansion of the lattice on glycolation.

It is probable that the mixed layer material is a mica/montmorillonite interstratification from the intermediate

position of the 001/001 peak between the 10Å spacing of the mica and the 12.4Å or 15.4Å (dependant on hydration) of montmorillonite. On glycolation the montmorillonite layers expand to 17Å and two resulting peaks are formed, an 001/001 combination of 10Å and 17Å which gives the main expanded peak at c. 12.5Å, and an 001/002 peak representing a combination of the 10Å mica and the 8.5Å 002 montmorillonite. This is the 'shoulder' on the main peak which ranges up from 9.2Å. An interstratification of this type also explains the expansion of the 5Å peak and strengthening of the 3.3Å peak on glycolation. Heating of such a mixed layer mineral should produce a contraction to somewhere between 10Å and the 9.8Å value of contracted montmorillonite. Some of the observed values lie in this range, but others are slightly greater than 10Å possibly due to the mica having a spacing slightly greater than 10Å, or the proportion of the montmorillonite being too small to cause a marked shift in the peak position.

Calculations were made in an effort to discover the relative proportions of mica and montmorillonite present. The broad nature of the peaks indicates that there is a range of composition present and that the interstratification is probably random in nature. The

data of Weaver (1956) indicate that there is from 20-35% of montmorillonite present, and those of MacEwan, Ruiz Amil and Brown (1961) indicate that from 20% to about 40% may be present.

Also present on the traces of the clay beds are peaks at 7.17\AA and 3.58\AA representing kaolinite. The peaks do not change on glycolation and are destroyed by heating.

ALKALI ANALYSIS of K-BENTONITES.

Samples of B1 (Loc 105 Edale), B2 (Loc 001), B5 and B6 (Loc 421) were analysed for K and Na using a flame photometer. The results are set out in the table below.

	%K ₂ O	% Na ₂ O
B1	2.30	0.87
B2	4.10	0.68
B5	4.17	0.42
B6	3.66	0.71

These results confirm the potassic nature of the K-bentonites. The percentage of potassium is similar to that found in K-bentonites from Kinnekulle, Sweden (Brystrom 1954; Hower and Mowatt 1966 p. 835), but is lower than the 7-4.5% usually found in K-bentonites with 20-40% expandable layers. This is undoubtedly due to the presence of kaolinite in the K-bentonites. The

K-bentonites analysed have similar proportions of expandable layers, and therefore the K₂O percentages should be similar. If the relative heights of the 001 kaolinite and 001 mica-montmorillonite peaks are compared it is found that as the proportion of kaolinite increases the K₂O% reduces. This is possibly due to leaching of the potassium at the time of formation of the kaolinite.

PETROGRAPHY OF K-BENTONITES.

Petrographic examination of thin sections of the K-bentonites reveals that they consist of extremely fine-grained, low birefringent clay material presumably the mica-montmorillonite, with a few scattered vermicules of kaolinite. Sections could only be obtained of K-bentonites that had been slightly calcified since collection of the others to retain their original texture proved impossible. In a section of B5 from Edale taken perpendicular to the bedding occasional small ragged quartz grains are present and a few carbonaceous streaks lie parallel to the bedding. Apart from these occasional streaks there is no evidence of any lamination within the bed.

CONCLUSION.

It is concluded that the clay beds all have an essentially similar clay mineralogy, and cannot be

differentiated one from the other with certainty on their clay mineralogy alone. They all contain a mixed layer mica-montmorillonite with 20-40% montmorillonite and variable amounts of kaolinite.

THE SHALE-MUDSTONES.

All the shale-mudstone samples were dark brown or black in colour and were collected from near the clay beds. Fig. 1.4c shown the positions from which the shale-mudstone samples were taken. Specimens 5, 7, 14A, 20 and 21 were all scanned from $3-30^{\circ}2\theta$ and specimens 9, 10, 12 and 13A were scanned from $4-15^{\circ}2\theta$ to cover the main clay mineral peaks and provide a check on the other results. The samples were treated in exactly the same way as those from the clay beds and the results are shown in Table 1.4e.

A typical trace of an untreated sample shows an asymmetrical peak at 10\AA indicating mica together with varying proportions of a mixed layer mineral, probably a mica-montmorillonite, that produces a shoulder or a peak on the high side of the mica peak from $10.5-11.3\text{\AA}$. There is also present a peak at 14\AA indicating chlorite, and a peak at 7.18\AA indicating kaolinite. The certain identification of both chlorite and kaolinite can be made since the 002 kaolinite and the 004 chlorite peaks

could be resolved from one another.

The glycolated traces tend to show a sharpening of the 10\AA and 14\AA peaks and a reduction of the mixed layer peak possibly due to irregular expansion having taken place. No regular expansion is seen as in the clay bed samples. Slight expansion of the 14\AA chlorite is occasionally observed on glycolation, possibly indicating the presence of a small proportion of expandable layers.

On heating the 14\AA peak remained or contracted to 13.8\AA a feature typical of chlorite. The mica peak sharpened at 10\AA and the mixed layer peak disappeared. The kaolinite peaks were also destroyed.

The clay minerals present in the shale-mudstone samples are considered to be chlorite, muscovite mica, a mixed layer mineral, probably a mica-montmorillonite and kaolinite.

This contrasts with the clay beds in the presence of chlorite and well crystallised mica and the absence of a strong mixed layer mica-montmorillonite peak at about 12\AA . Typical traces of specimen 14A are illustrated (Fig. 1.4i).

PETROGRAPHY OF SHALE-MUDSTONES ASSOCIATED WITH THE K-BENTONITES.

Sections taken perpendicular to the bedding of the shale-mudstones illustrate clearly their detrital nature

and marked difference from the K-bentonites. Irregular carbonaceous laminae are frequent and usually rather opaque. Most of the shale-mudstone consists of fine grained flakes of clay minerals or mica less than .005cm. long. Some flakes of highly birefringent muscovite up to .01cm. long are clearly detrital and are thought to represent the 2M high temperature material identified on the diffractometer traces. Quartz grains rarely exceed .005cm. in diameter and are mostly very small and form part of the general ground mass. Chlorite could not be identified in thin section, a clear indication of the inadequacy of thin section examination of such fine grained rocks. The shale-mudstones present a clearly detrital aspect in contrast to the K-bentonites.

ORIGIN OF THE CLAY MINERALS.

Several other tests were tried in an attempt to find the probable source of origin of the mixed layer mica-montmorillonite found in the clay beds and the shale-mudstones. From the diffractometer traces these mixed layer minerals appear to be different because of their different positions and reactions on glycolation.

Clay bed samples 6 and 18 and shale-mudstone samples 14A and 23 were subjected to potassium saturation with 1N KOH. After soaking for fifteen hours in 1N KOH and then heating to 90°C for one hour in 1N KOH, the clay bed

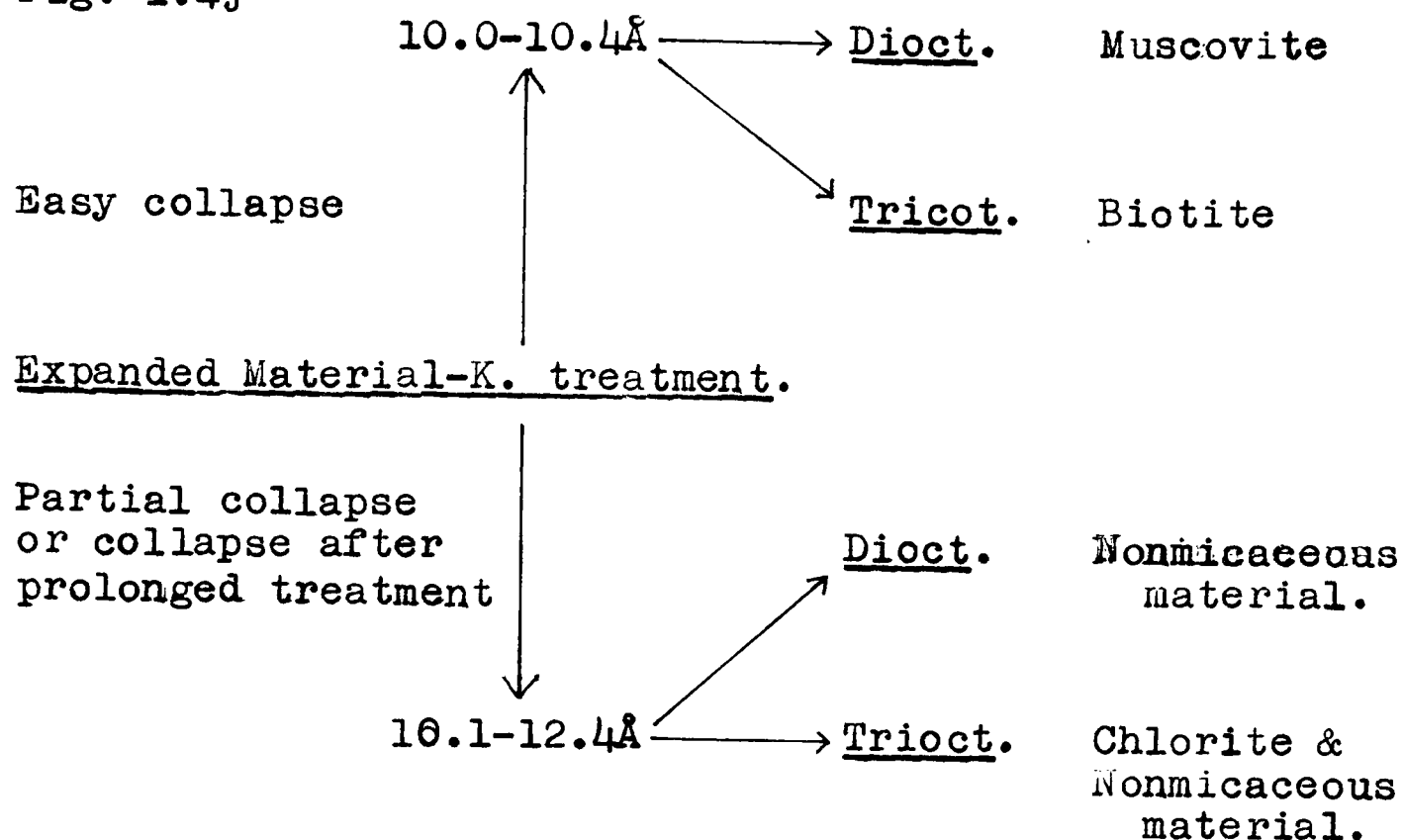
samples collapsed to spacings of 11.04\AA and 10.84\AA . This only partial collapse of the expansible layers suggests that the mixed layer clay mineral was derived from a non-micaceous source material (Weaver 1958) with a low interlayer charge (less than 150 meq. per 100 gm. ignited material, Weaver 1958).

The shale mudstone samples were soaked in 1N KOH and after only four hours the majority of the mixed layer material had collapsed to 10\AA which indicates that it was derived from micaceous material with a higher interlayer charge (greater than 260 meq. per 100 gm. ignited material). After twenty four hours no further contraction had taken place.

Hower and Mowatt (1966) have shown that the mica-montmorillonites form a distinct mineralogical group, and that the proportion of expandable layers present in the structure is related to the total charge and also to the Cation Exchange Capacity (C.E.C.) Their work tends to confirm that the K-bentonites were derived from material of lower interlayer charge than that from which the mixed layer material in the shale-mudstones came.

Fig.(1.4j) summarises the effects of KOH treatment and the probable origin of the material.

Fig. 1.4j



Adapted from Weaver 1958.

Nonmicaceous material = generally volcanic and mafic rocks and minerals.

The clay bed material was determined as being dioctahedral with 060 at 1.498\AA , this was checked on specimens 6 and 18.

With the shale-mudstones it could not be certainly proved that the mica is dioctahedral from the diffractometer trace since both dioctahedral and trioctahedral minerals are present, however it is probable that a small peak at 1.50\AA represents dioctahedral mica.

Yoder and Eugster (1955) demonstrated that the different polymorphs of mica can be distinguished from their diffractometer traces. They also found that the 2M structure is typical of high temperature muscovites which in the sedimentary environment are typical of a low temperature of formation.

The diagnostic reflections occur in the region $20-35^{\circ} 2\theta$ and it is unfortunate that peaks of quartz, kaolinite and chlorite all occur in this range. In several traces of the shale-mudstone specimens peaks occurred in the region of 3.21\AA and 3.00\AA which are diagnostic of the 2M muscovite structure of Yoder and Eugster. Thus this mica in the shale-mudstones is considered to be of detrital origin.

The broad nature of the peaks produced by randomly oriented mounts of the clay bed samples in the range $20-35^{\circ} 2\theta$ indicate that the mineral probably has a 1Md structure. The generally diffuse nature of the peaks due to the mixed layering and the interference of the 002 kaolinite peak makes certain identification of the diagnostic $11\bar{2}$ and $11\bar{2}$ peaks difficult. Peaks typical of the 2M structure are not present. It is concluded that the clay beds contain material formed at a low temperature.

Kaolinite occurs in both the clay beds and in the shale-mudstones. It could be formed by weathering of

micaceous material in recent times, or, in the case of the shale-mudstones, in the source area before deposition of the material. Poncelet and Brindley (1967) have shown that montmorillonite can also alter to kaolinite under acid conditions at low temperatures and suggest that acid ground water with a pH of 6.5-5.0 can cause the change. Chlorite is restricted to the shale-mudstones and is probably of detrital origin.

The origins of the clay minerals in the clay beds and in the shale-mudstones can be summarised as follows.

CLAY BEDS. - THE K-BENTONITES.

Mica/montmorillonite - Derived from non-micaceous material with low interlayer charge at low temperatures. 1Md structure.

Kaolinite - Alteration of the mica/montmorillonite possibly by acid ground water.

SHALE-MUDSTONES.

Mica - Mainly detrital, 2M structure, high temperature of formation.

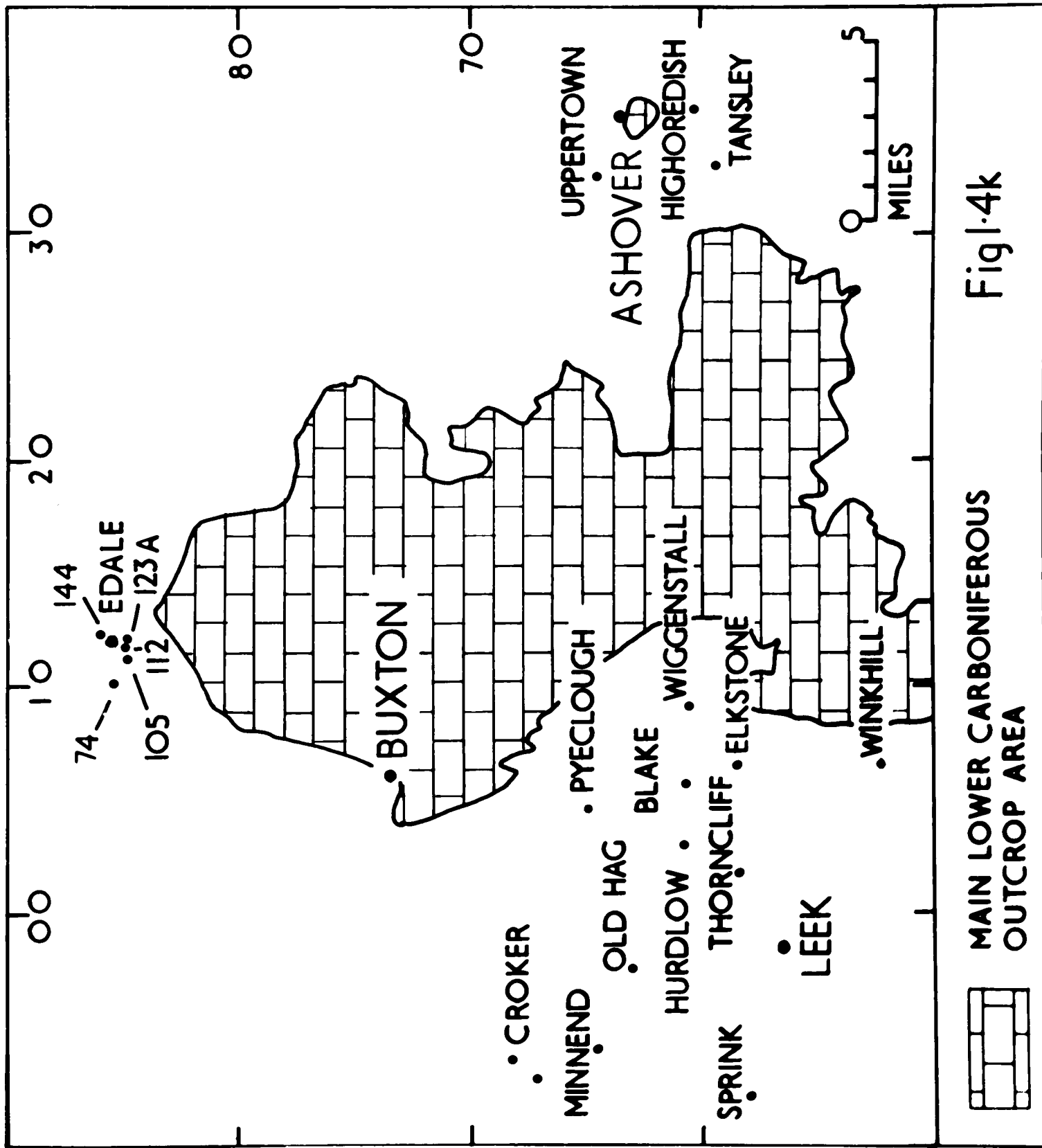
Mica/montmorillonite - Derived from mica by weathering.

Chlorite - Detrital.

Kaolinite - Possibly partly detrital from original source.

ORIGIN OF THE CLAY BEDS.

It is clear from the above evidence that the shale-



mudstone and the clay beds are of totally different origin. This is also born out by the field characteristics of the two lithofacies. The shale-mudstone is obviously the 'normal' bottom sediment of the area containing detrital quartz, 2M mica and chlorite. The clay beds represent abnormal events in the basin area, and the bulk of material produced is insignificant compared with that of the shale-mudstone.

The occurrence of the clay in thin beds of such great lateral extent (see distribution map, Fig. 4k) makes it impossible that they were deposited by ordinary bottom currents or by turbidity currents, as also does their totally abnormal mineralogy of a mica-montmorillonite derived from a non-micaceous source at a low temperature.

The only real possibility for their origin is that they are the altered and somewhat weathered products of showers of volcanic ash. The field characteristics of great lateral extent, distinctive mineralogy, and sharp contacts with the shale-mudstone support this view.

The mineralogy can be simply explained, since the fine grained glassy material of an ash shower would alter easily to montmorillonite. In the marine environment some of the montmorillonite layers would fix potassium and alter towards a mica structure, so producing the mixed layer structure. The kaolinite

present could be derived from the weathering of original feldspar or by alteration of the mixed layer material under acid conditions. Grim (1953 p.362) records that up to 50% kaolinite can occur in bentonites.

This simple explanation of the conversion of montmorillonite to a mixed layer structure by potassium fixation is probably a simplification of the truth. Hower and Mowatt (1966) have shown that mixed layer structures with only 20% expandable layers have structural charges similar to illites and cannot simply be montmorillonites which have absorbed potassium as the interlayer cation. They consider that compositional changes such as lateration of the tetrahedral Al-Si ratio are also necessary. This theory is supported by Beall and Ojakangas (1967) who consider that Al^{3+} has replaced Si^{4+} in the tetrahedral layer of a mixed layer mica/montmorillonite from a Cambrian K-bentonite in Missouri.

It is concluded that the clay beds are potassium bentonites, the alteration product of ash showers of volcanic origin.

COMPARISON WITH DESCRIBED K-BENTONITES.

Many K-bentonites have been described from the United States and all correspond well with the present material. Mosler and Hayes (1966) described eight K-bentonites from the Ordovician of Iowa, and found them all to be randomly interstratified mica/montmorillonite

with from 23-30% montmorillonite. The 001/001 spacing of their material was 11.5Å, on glycolation peaks of 12.6Å (001/001) and 9.7Å (001/002) appeared. Heating caused a collapse to 10.1Å, and potassium treatment identical to that used in the present study gave a contraction to 10.7Å. They also showed that the material had a 1Md structure. Thus this material is very similar to the K-bentonites described here, the only difference being the absence of kaolinite in the Iowa material. The field characteristics are identical with those of the beds in the present area, and Mossler and Hayes conclude that the Iowa K-bentonites are of volcanic origin.

Weaver (1953) described K-bentonites from Pennsylvania with 20-25% montmorillonite and Huff (1963) found a range of 20-40% expandible layers in K-bentonites from Kentucky. These occurrences both agree with the present results.

There are very few records in the literature of bentonites within the British succession. The best known example is probably the 'Green Streak' which occurs in the Stockdale Shales of the Lake District, and appears to be very widespread.

Butler (1937) in recording the log of a boring through the Silurian at Walsall records 51 separate horizons of green to mauve bentonite beds ranging from

.06" to 7" in thickness, of which only 16 are 1 inch thick or greater. The beds are concentrated mainly at the top of the Upper Llandovery and the base of the Wenlock Shale. Butler records that many beds are rich in biotite of volcanic origin but none has been seen in the beds at present under discussion probably due to their weathered nature. The clay minerals in the rocks from the Walsall bore could not be identified in thin section due to their small grain size.

The reaction of the clay when placed in water is described by Butler. The clay first expands and flakes, then "falls into a fine flour-like aggregate." Similar green clay beds found in the Wenlock Limestone by the author at Coats Quarry on Wenlock Edge react in the same way when placed in water as also does a similar specimen from The Wrens Nest, Dudley (also Wenlock Limestone).

X-Ray diffraction studies on these beds from the Wenlock Limestone confirm that the clay mineral present is a mixed layer mica-montmorillonite with about 25-35% montmorillonite. K-saturation gives incomplete collapse of the lattice and the mica has a 1Md structure. Thus this material is very similar to that present in the Namurian K-bentonites described here. In the case of the bentonite from Wenlock Edge it is interesting to note that the calcareous shales that enclose it contain detrital

mica and chlorite, clearly of different origin to the K-bentonite clay.

Thus the K-bentonite material in the present area is typical of K-bentonites already shown or suspected to be of volcanic origin.

CORRELATION WITH OTHER AREAS.

EDALE, DERBYSHIRE.

The exposures of the Namurian in Edale were described by Hudson and Cotton (1945). The correlations of the faunal horizons have been described in the previous section. (see also Fig.1.1b).

In order to confirm the correlation of the "E.aff. pseudobilingue" band of Alport (Hudson and Cotton 1943) and Edale with the E. grassingtonense band of north Staffordshire, the exposures in Edale were visited to see if the B2 pyritic bentonite was present beneath this horizon as in north Staffordshire.

At locality 105 of Hudson and Cotton (1945) (H.C.105) a 0.5-0.7 cm. pyritic clay bed occurs 70 cm. below the faunal horizon, thus correlating well with the position of B2 in north Staffordshire below the E. grassingtonense horizon. Also present at this locality is a 1.0 cm. yellow clay bed 164 cm. below B2. This horizon has not been seen in north Staffordshire since it would occur in the topmost part of the B unit of protoquartzites and was probably eroded by turbidity

current activity. This is the lowest K-bentonite horizon found and is termed B1, it is similar in appearance to B5 and B6 in not being pyritic. At Loc. H.C. 74 the "E. aff. pseudobilingue" horizon is again seen, and the two bentonites B1 and B2 are present in almost exactly the same positions relative to each other and to the faunal horizon as at Loc. H.C. 105. Thus the evidence of the K-bentonites confirms the correlation of the "E. aff. pseudobilingue" horizon of Hudson and Cotton with the E. grassingtonense horizon of north Staffordshire.

At Loc. H.C. 112 a good exposure of shales in a river bluff and in a small tributary stream has revealed the presence of beds B3, B5 and B6 in the usual positions relative to the Ct. edalensis and L. longirostris faunal horizons. At this locality B5 is 4 cm. thick and is the thickest K-bentonite found in the region. These three beds are not pyritic and appear as greenish clay or mudstone weathering to yellow or orange.

Bed B3 is underlain by two semi-continuous siderite beds 2 cm. and 3 cm. thick and is the only K-bentonite horizon associated with siderite beds in this way. This horizon is also seen in a shale scar about 100 yds. upstream from Loc. H.C. 123a, where it is also associated with siderites as at H.C. 112.

K-bentonites B5 and B6 are exposed at the top of

the cliff at H.C. 112 and also in the tributary stream where Ct. edalensis was found about 400 cm. above the K-bentonites. These two horizons have also been found about 20 yards downstream, from Loc. H.C. 144 which contains abundant Ct. edalensis. Slight faulting and crumpling of the shale mudstone makes the exact distance of B5 and B6 below the Ct. edalensis band impossible to determine.

Upstream from Loc. H.C. 144 near Loc. H.C. 145 a small tributary enters from the west just before a footbridge is reached. At the point where the tributary enters the main stream a 1.5 cm. pyritic clay bed is exposed. This horizon is about 5 metres above the Ct. edalensis horizon and is correlated with bed B7 of north Staffordshire.

In Edale all the K-bentonites seen in north Staffordshire have been found with the exception of B4, and this bed is known to be discontinuous in north Staffordshire. Only one additional K-bentonite is present in Edale (B1) and its absence in north Staffordshire is easily explained. Horizons B2, B3, B5, and B6 have all been found at two different localities in Edale and B7 at one locality.

Specimens of B1 (Loc. H.C. 112) were examined qualitatively using an X-ray diffractometer exactly as described for the Staffordshire specimens. Traces

of these beds were found to be very similar to those from their equivalent beds in Staffordshire in the cases of B5 and B6, and the bentonitic nature of B1 was confirmed. All three beds contained kaolinite together with a mixed layer mica-montmorillonite. The traces showed broad peaks indicating random mixed layer mica-montmorillonite with an 001/001 spacing ranging from 12.25Å-12.46Å at maximum peak height. On glycolation expansion occurred to 12.92Å-13.47Å 001/001 with 002/001 at about 9-9.5Å. Heating to 550°C for 2 hours caused collapse of the lattice to near 10Å in each case. Thus the material making these beds can be seen to be virtually identical to that forming the K-bentonites to the south in Staffordshire (Compare with Fig. 1.4d.)

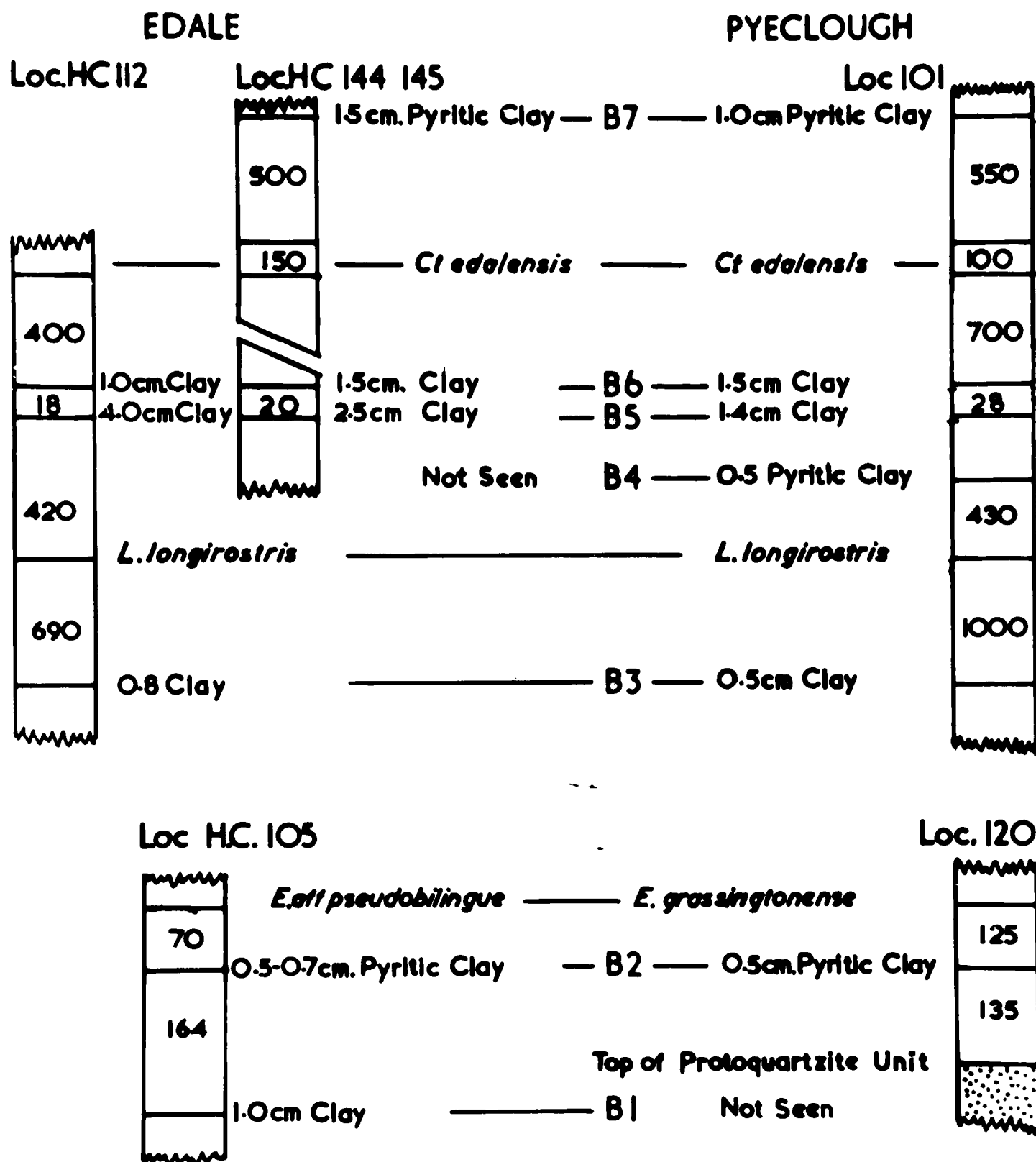
Fig. 1.41 shows the relative positions and thicknesses of the K-bentonites in the Edale exposures and a comparison with the Pyeclough section of north Staffordshire.

ASHOVER, DERBYSHIRE.

Ramsbottom, Rhys and Smith (1962) in their paper on the Ashover boreholes recognised certain horizons that could be correlated between the three boreholes. These horizons include pyrite beds and green to brown mudstones. They state (p.118) that: "It is found that 'cank' bands, pyrite horizons and light coloured partings

Fig. 41

CORRELATION OF K-BENTONITES BETWEEN LOCALITIES
IN EDALE AND N. STAFFS. (PYECLOUGH SECTION)



can often be correlated with certainty; for instance, in the E. bisulcatum and C. cowlingsense Zone, $7\frac{1}{2}$ inches to $10\frac{1}{2}$ inches above the Leiopteria longirostris horizon, which is in each case a 'cank', a $\frac{1}{4}$ inch band of pyrite is present in all three holes; near the top of the zone are two partings of green or greenish-grey mudstone varying in thickness from $\frac{1}{2}$ inch down to a coloured bedding plane." These beds correlate well with the beds B4, B5 and B6 in north Staffordshire in their positions relative to the faunal bands.

Also mentioned in the borehole logs are other pyrite or brown mudstone horizons which appear to correlate with one another. In the Tansley and Uppertown boreholes a $\frac{1}{4}$ inch pyrite bed is recorded above the Ct. edalensis horizon, in each case associated with a fauna. This bed can be correlated with B7 in the basin where it is also pyritic.

In each borehole there is a brown mudstone horizon below the L. longirostris horizon, which could correlate with B3. A half inch brown mudstone and 1 inch pyritic bed from the Tansley and Highoredish bores may correlate with one another but their equivalent has not been seen in the basin area.

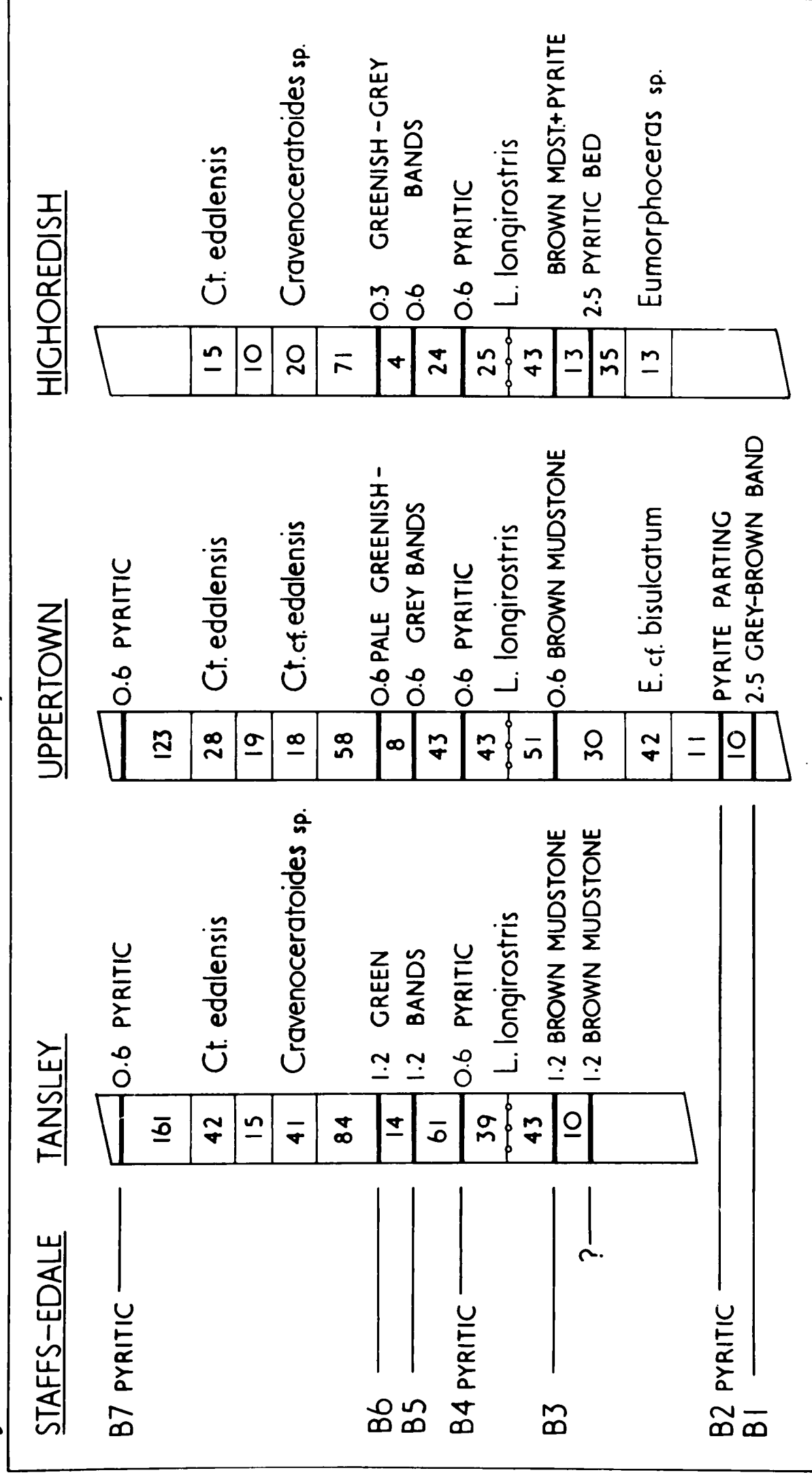
Below the horizon containing E. cf. bisulcatum and Cravenoceras sp. in the Uppertown bore there is a

pyrite parting and below that a 1 inch band of grey-brown mudstone. The pyrite horizon can be correlated with the pyritic bed B2 of north Staffordshire and Edale, and the grey-brown mudstone is possibly the equivalent of B1 in Edale. These horizons are not present in the other bores. If these correlations are correct the bed with E. cf. bisulcatum and C. sp. could represent either of the E. erinense-E.ferrimontanum or E. grassingtonense horizons of the north Staffordshire basin, or less probably, both of these horizons merged into one marine horizon.

Fig. 1.4m shows the suggested correlations between the three Ashover boreholes, based on the detailed logs of Ramsbottom, Rhys and Smith (1962). Correlations of these beds with those found in north Staffordshire and Edale are also shown. Through the courtesy of Dr. W.H.C. Ramsbottom some specimens from the Ashover boreholes were examined and samples prepared for X-ray diffraction.

The only sample still preserved of the 'green-grey mudstone' bands used for correlation by Ramsbottom et al. (1962) is from 246ft. 10in. in the Highoredish borehole which was recorded as a parting in the log but the specimen shows it to have been at least 3mm. thick. In appearance it is very similar to material from the

Fig 1.4m



Correlation of K-bentonites B1-B7 with the Ashover boreholes. Based on the logs of Ramsbottom et al. (1962). Not to scale, thicknesses all indicated in cms.

basin areas and diffractometer traces showed that its composition is nearly identical. The untreated sample gave a broad mixed layer mica-montmorillonite peak at 11.25\AA which on glycolation expanded to 12.1\AA and reduced to 10\AA when heated. This behaviour being identical with that observed from material from the basin area with the minor difference that the 11.25\AA untreated spacing is slightly lower than that observed for the equivalent bed in the basin areas ($11.9 - 12.5\text{\AA}$). This may be due to a smaller proportion of water layers in the montmorillonite material from the boreholes.

A pyritic horizon from 251ft. in the same borehole, tentatively correlated with B3 in the basin area was found to be rather poor in clay material, the reflections being very weak probably due to the concentration of pyrite in the bed. However weak reflections from mixed layer clay material were obtained that were affected by glycolation and heating. Kaolinite was also present.

The shale mudstones associated with these beds yielded the same clay minerals as those found in the basin areas, but the clay material was much more degraded a feature consistent with the slow rate of deposition on the block area, leaving any deposited clays in contact with sea water for a long time. The

mica present was found to be altered to mixed layer material as was the case in some of the shales examined from the basin area. Chlorite was present in a sample from adjacent to the green mudstone in the Highoredish borehole, but only doubtfully present in a specimen from adjacent to the pyritic horizon examined. Quartz was present in the shale-mudstone samples but not in the bentonites, and kaolinite was present in shale-mudstone and bentonites alike.

The presence of kaolinite in the borehole specimens seems to indicate that it is not derived from a mica-montmorillonite by recent weathering but that it is of earlier origin, perhaps being derived from the alteration of feldspathic material. However the presence of kaolinite in the normal shale-mudstones indicates that it was probably stable in the Namurian marine environment possibly due to the chemical conditions prevailing at and below the sediment/water interface. Poncelet and Brindley(1967) consider that acid ground water could alter montmorillonite to kaolinite and this may have happened in the case, alternatively acid conditions prevailing below the sediment/water interface could have effected the change.

* * * * *

SEDIMENTOLOGY INTRODUCTION.

In the following sedimentological treatment of the Upper Pendleian and Lower Arnsbergian sediments, emphasis is laid on the sedimentary structures present and interpretation of the modes of transport and deposition of the various lithofacies recognised. Current direction data and areal variations are described where appropriate.

Detailed geochemical studies have not been performed on the sediments but the clay minerals present have been determined by whole rock analysis using a Siemens X-ray diffractometer.

Extensive development of secondary minerals including quartz and siderite in the protoquartzites, and quartz with ankerite, calcite-dolomite and pyrite in the calcareous siltstones makes worthwhile grain size and shape analysis impossible as is pointed out in the petrography sections.

PART 2: SEDIMENTOLOGY.

SECTION 2.1

THE PROTOQUARTZITES.

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2.1A

BRIEF DIAGNOSIS.

The protoquartzites generally consist of parallel sided beds set in shale-mudstone. They have sharp bases and gradational or sharp tops. Scour, tool and load structures are of frequent occurrence on bases of beds. Bioheiroglyphs are also common on the bases of beds. Internally the beds may be graded and frequently are laminated or rippled. Mineralogically the beds are protoquartzites, K-feldspar is absent and grains are frequently silica cemented. Siderite may replace parts of beds and occurs as nodules in the shales associated with the protoquartzites. Calcite is absent.

2.1B

STRATIGRAPHICAL POSITION.

Three units of protoquartzites set in shale-mudstone occur in the succession studied and their positions relative to the faunal bands have been shown in Fig.1.1a. The three units are termed A B and C, the lowest in the succession being A.

2.1C

GENERAL FIELD APPEARANCE.

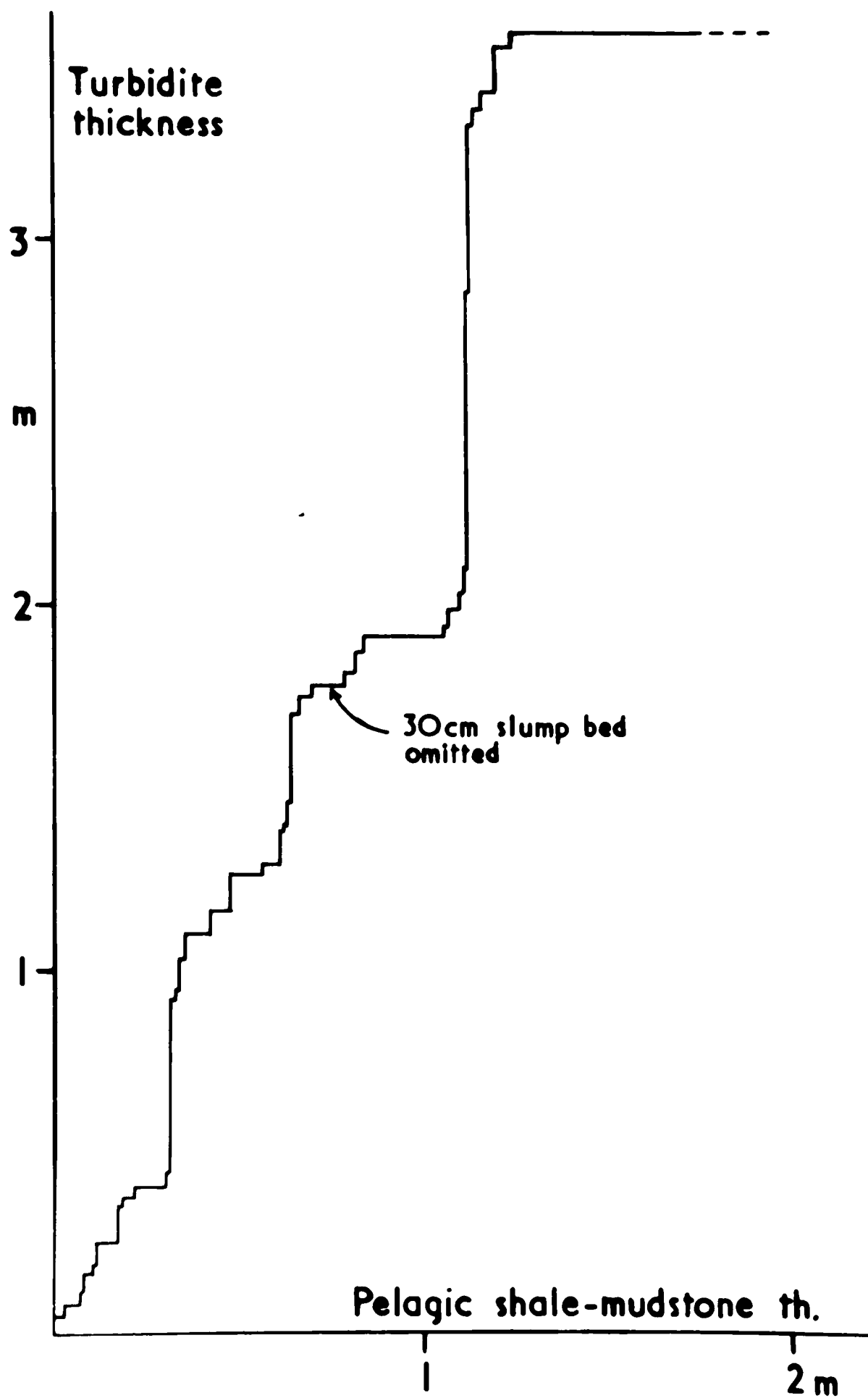
The above diagnosis takes no account of the thicknesses of the individual sandstone beds or of the proportions of sandstone and shale-mudstone in any one section. There is in fact a continuous variation from sections consisting of nearly 100% sandstone and no shale-mudstone, to sections consisting entirely of shale-mudstone with siderite nodules. There is therefore a gradation from sections consisting of 'protoquartzites with shale-mudstone and siderite beds and nodules' to sections purely of 'shale-mudstone with siderite'. When protoquartzites are present in a unit it will be referred to as being of protoquartzite lithofacies. The variations in proportion of sandstone and shale-mudstone in measured sections are considered in section 2.1K.

In any one area sand beds are not evenly distributed through a unit of this lithofacies. There tend to be groups of thick sand beds separated by thin beds of shale-mudstone, alternating with groups of thin sand beds separated by thicker shale-mudstone horizons. This feature is illustrated by the accumulative sandstone/shale-mudstone diagram (Fig.2.1Ca) which seems to indicate that there is a certain minor cyclicity about the deposition of the sand beds. The lack of long sections suitable for measuring makes further study of this interesting

Fig 2.1Ca

Cumulative turbidite/shale-mudstone diagram

Minor Cycles Unit A Protoquartzites Loc O12



feature impossible. Since there is a continuous variation in the ratio of sandstone to shale-mudstone no attempt is made to institute defined subdivisions in the series. Instead, typical sections in the proto-quartzites will be described which illustrate all the typical features of the lithofacies.

Within the area studied there is a general tendency for individual sandstone beds, and for units of proto-quartzite development to decrease in thickness from the south to the north and north-east - this being the general direction of sand transport as will be shown later. Sections in the south of the area at Sprink, Shirkley, and near Endon Mill show thick developments of protoquartzites here termed of Proximal type, whilst towards the Limestone Massif area (e.g. Pyecclough) where sandstone beds are thin and separated by thicknesses of shale-mudstone the developments are termed Distal. Intermediate types of development are most frequent in the area and can be seen in the Hurdlow, Thorncliff, Gun Hill and Croker Hill areas.

Logs of typical proximal and distal sections are given here (Figs.2.1Cb-d^{*}) to illustrate the range of appearance of the protoquartzites. In the field area exposure is generally poor and few sections are available where long unbroken sequences are exposed. Thus the logs measured by the author at any one locality probably

* Figures in folder.

do not reflect the whole variety of sediment to be found in that area. Statistical work on the available measured sections is largely meaningless because of this and thus no advanced statistical treatments have been presented. The internal structures of the sand beds cannot be easily seen in the field as the laminations do not weather out and specimens generally had to be cut to show the internal structures. Laminations in some beds however are picked out by plant fragments and the beds readily split along these laminations.

The protoquartzite beds always have sharp bases and the bases exhibit a variety of structures which are described below (Section 2.1D). The tops of the beds may appear sharp and rippled or there may be a gradation from protoquartzite at the base to black shale-mudstone via green silty shale or green shale-mudstone. The significance of the beds with gradational and sharp tops will be discussed later.

Brief statement of the main features of Proximal and Distal sequences.

Proximal.

Proximal sequences contain a high proportion of protoquartzite beds greater than 20cm. thick which are usually massive, and may have a thin graded top. Beds are parallel sided and tops and bottoms of successive

beds may become welded together. Bottom structures are usually poor but load casts, flute marks and tool marks occur. Pebbles are locally concentrated in load pockets and may occur within beds. Thickness of shale-mudstone between beds rarely exceeds a few centimetres.

Distal.

Beds are parallel sided and have sharp bases which frequently show tool markings and unroofed burrows. The beds are normally less than 20cm. thick and usually have tops which grade to silty green shale and then to black shale-mudstone. Internal lamination of the beds is frequent and ripples may also occur.

Having briefly indicated the range of variation to be found within the protoquartzites, the individual structures which occur either as sole markings or within the bed will now be described.

2.1DEXTERNAL SEDIMENTARY STRUCTURES OF THE PROTOQUARTZITES.

SOLE MARKINGS.

Introduction.

Classification of sole structures can theoretically be made by two methods. One being a purely morphological description of the structure without any implication of the mode of formation. The name applied to the structure must then carry no suggestion as to how it was formed. The second method is a classification according to the origin of the structure, and in this case the name applied to the structure has a genetic significance. In practice a mixture of the two methods is employed, those markings whose origin is definitely known generally have genetic names whilst those where there is doubt as to their origin generally are given morphological names.

The most useful classification to a sedimentologist is a genetic one, in that structures produced by the same processes can be grouped together. The best classification produced on these lines is the one given by Dzulynski and Walton (1965). This classification distinguishes three categories of marks produced on the sole of the bed by the passage of a current.

These are:-

- a) Scour marks - produced by the current acting alone and eroding the bed as it passes.
- b) Tool marks - produced by an object moving with the current.
- c) Deformation marks caused by currents and not directly related to scour or tool marks - produced by the drag effect of a current passing over the floor and deforming the mud.

Groups a and b produce quite easily definable groups of markings, but group c is more difficult to define as deformation of the mud can take place at all times from predeposition through syndeposition to postdeposition, and this often means that structures are produced both within the sand bed as well as in the underlying mud. This is so because the 'drag effect' can act through a thickness of deposited sand and still affect the underlying mud, this drag being caused by the current flow.

It would be best to limit group c to predepositional deformation marks produced by the current. This seems to be the sense in which Dzulynski and Sanders (1962) intended the group and it is expressed this way in Dzulynski and Walton (1965) as it includes transverse wrinkles and the substrate cased by the passage of the current. Mud ridges and tension cracks associated with

tool marks are also included, but none of these were seen in the area.

SCOUR MARKINGS.

Bottom structures produced by scour are generally rare in the area studied. They have been seen most frequently in the sections showing proximal type beds at Sprink, Shirkley and near Endon Mill. They also occur in sections at Hurdlow and Croker Hill associated with strong developments of protoquartzites.

Flute moulds.

Flute moulds have received a great deal of attention from sedimentologists because of their frequent occurrence and variety of shape (e.g. Rücklin 1938, Vassoevic 1953, Crowell 1955, Ten Haaf 1959 and Mc Bride 1962). Dzulynski and Walton (1965) and others have described and figured various forms of flute moulds, and it is generally accepted that their origin is due to scour of the substrate by eddies in the main current. Thus they form during the turbulent passage of a current over the substrate or by local turbulence created by irregularities in the substrate.

It is known from experiments (Dzulynski and Simpson 1966) that the presence of tools in a current greatly facilitates the production of scour marks. The scour marks may be formed by eddies round a tool moving near

the bed or by modification of original tool marks by scouring.

In the present area flute marks are generally scarce, and are most often found in proximal type sequences as at Sprink where one bed was observed with bulbous flute moulds and another with scattered flat flute moulds. At Shirkley a single small example of a terraced flute mould (Fig.2.1Da) was found where scour had taken place into a laminated sand layer. Away from the proximal area of sand deposition flute marks are occasionally found but usually are associated with the thicker sand beds in sequences in which the amount of sand deposition was much greater than that of shale-mudstone.

Since flute moulds are rather uncommon in the present area no detailed work has been done on them. It is however of interest that they have only been found in the more proximal areas of sand supply to the basin or in areas of strong sand supply within the basin area.

Longitudinal ridge structure.

This structure has been observed on beds near Dalehouse Wood 060 and at Shirkley 050. The ridges appear to be the infilling of grooves cut by scour and are therefore predepositional as has been suggested by Craig and Walton (1962), Dzulynski and Walton (1965). The bulbous end of the ridge is interpreted as being at the upcurrent end as is consistent with an origin due to

scour, and this interpretation is confirmed by prod marks that occur in conjunction with the longitudinal ridges. The ridges may break up into a 'scaly structure' as in Fig.2.1Db. This is very similar to that figured by Dzulynski and Walton (1965), Figs.45A, B and by Craig and Walton (1962).

In size the ridges range from 1mm. to 10mm. wide with a relief of 0.5mm. to 2mm. Patches of longitudinal ridges may be present on the base of a bed possibly due to disturbances around an object carried close to the substrate by the current (see Fig.2.1Dc). This feature was also observed by Dzulynski and Walton (1965, Fig.73).

The formation of both longitudinal ridges and the scaly pattern into which they often grade has been explained by Dzulynski (1966) in terms of a convection motion within the sand settling on the mud. Where an upcurrent exists in the settling media a ridge is pulled up from the floor and at a down current scour takes place on the floor giving a hollow. Under pure gravity settling a polygonal pattern of ridges and hollows is obtained. When the settling is also under the influence of horizontal motion then longitudinal ridges and hollows are produced (Ibid. Fig.9a p.13). The ridges mark the lines of adjacent ascending currents and the hollows the descending currents.

It is therefore likely that the ridge structure and its associated scaly pattern are syndepositional structures due to convectional settling of the sediment.

Structures which are at least in part synonymous with 'longitudinal ridges' are 'load cast striations', 'syndromous load casts' and 'squamiform load casts' Ten Haaf (1959); 'elongate flute casts' Keunen (1957); 'longitudinal ripple marks' Kelling (1958); 'longitudinal ripple load casts' Kelling and Walton (1957), and 'sludge casts' Wood and Smith (1959).

Fleur-de-lys moulds. Fig.2.1Dd

One fine example of an isolated fleur-de-lys mould (Dzulynski and Walton 1965 p.61) was found at Shirkley associated with strong prod moulds and some longitudinal ridge moulds. The structure is clearly allied to the 'longitudinal ridge structure' and possibly marks the starting point for a series of longitudinal ridges that die out in a down stream direction. Thus it probably marks the point of initiation of a number of convection cells of the type that Dzulynski (1966) considers to have formed the longitudinal ridge structure.

Scour structure allied to longitudinal ridges.

Fig.2.1Dda.

One specimen from Shirkley showed a patch of

small impersistent longitudinal ridges. The patch is elongated in the direction of current flow and narrows in the downcurrent direction until it dies out. This structure occurs on the base of a bed which is generally smooth apart from small prod moulds, and the structure possibly represents scour marks behind an object fixed in the substrate over which the current was flowing. It may also have been formed just after deposition during convective settling of the sand whilst still under the influence of the horizontal flow of the current.

Scour behind pebbles.

A few beds of the proximal type contain pebbles of quartz and sometimes these became fixed in the substrate while the main bulk of the current continued to flow. The flow of the current caused scouring to take place around the pebbles (Fig.2.1Ddb). The pebbles, which were originally trapped in the muddy substrate, are now cemented in the base of the sandstone bed and protrude from the base of the bed.

One bed also showed irregular scour marks behind the former positions of large mud flakes (see Fig.2.1Ddc). Identical scour structures around pebbles have been figured by Ksiazkiewicz (1961) from the Carpathian flysch. He also observed similar structures formed around wood and sandstone fragments.

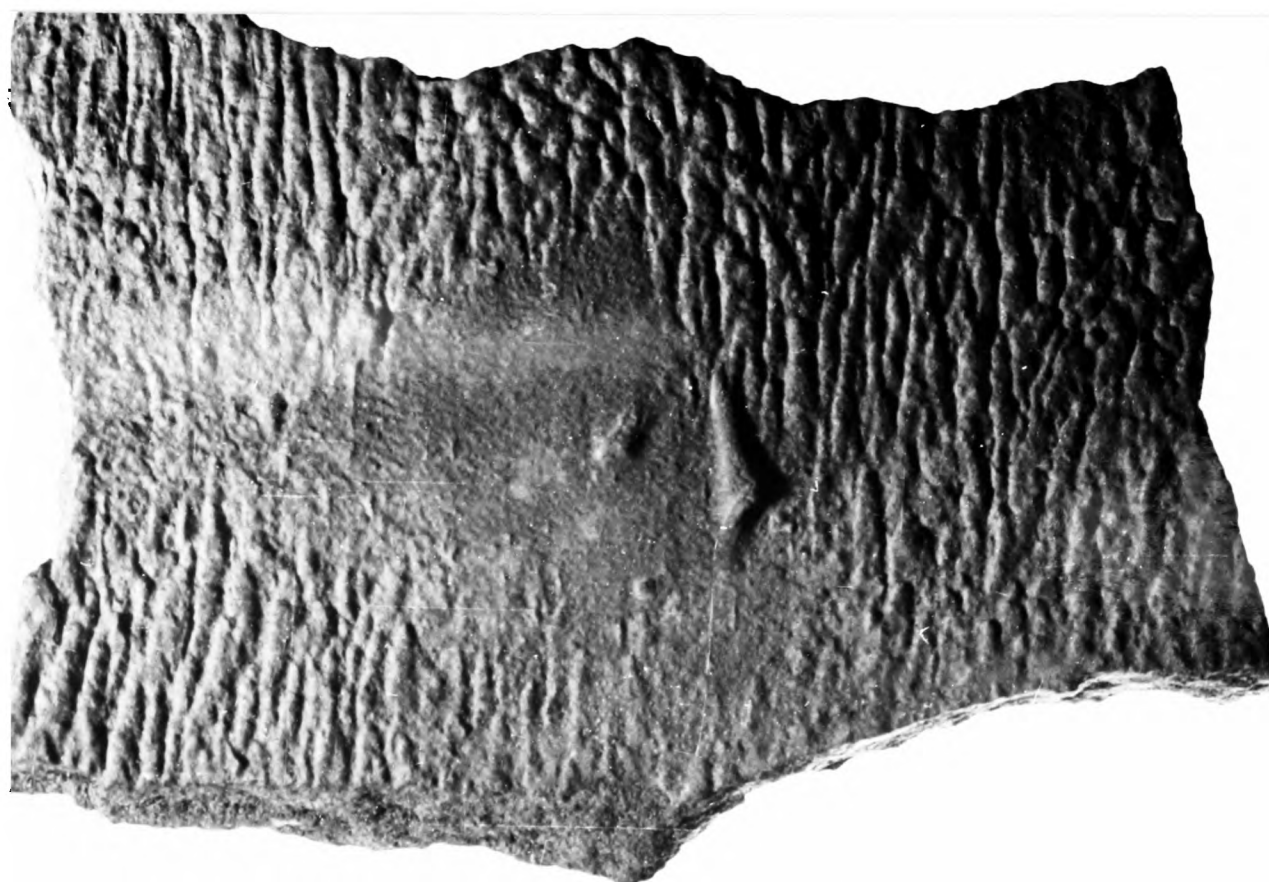
GENERAL CONSIDERATION OF SCOUR MARKS.

All the features indicative of current scour described here are rare in the present area so only a limited amount of information can be gained from them. The most important point is that they were found mainly in the area interpreted as being proximal to the source of the currents supplying the sand material. From the few examples seen there does not seem to be any preferred thickness of bed associated with scour structures. The thicknesses of the beds having these structures range from a few centimetres up to 50cm. in thickness, but it does seem that flute marks are strongly associated with sequences of thicker sand beds rather than with sequences where sand beds are usually less than 10cm. thick. Thus they could be associated with the thicker or denser sediment laden currents. Within the basin area their association with stronger developments of protoquartzites also supports this theory.



Fig. 2.1Da

Small terraced flute mould. Unit C. Loc.054.



CENTIMETRES

5

10

Fig. 2.1Db

Longitudinal ridge structure breaking up into scaly pattern, with prod mould near centre. Loc.064.

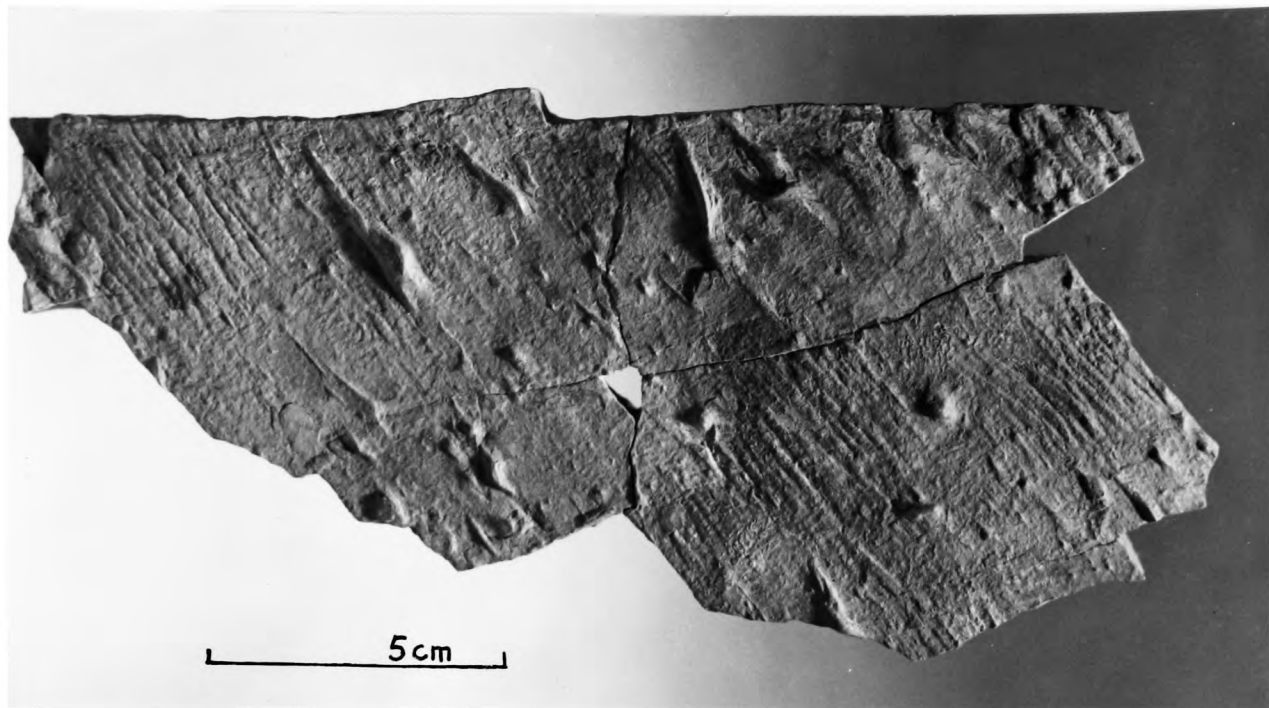


Fig. 2.1Dc

Patches of l-ridges associated with prod moulds on base of protoquartzite bed. Unit C. Loc.054.

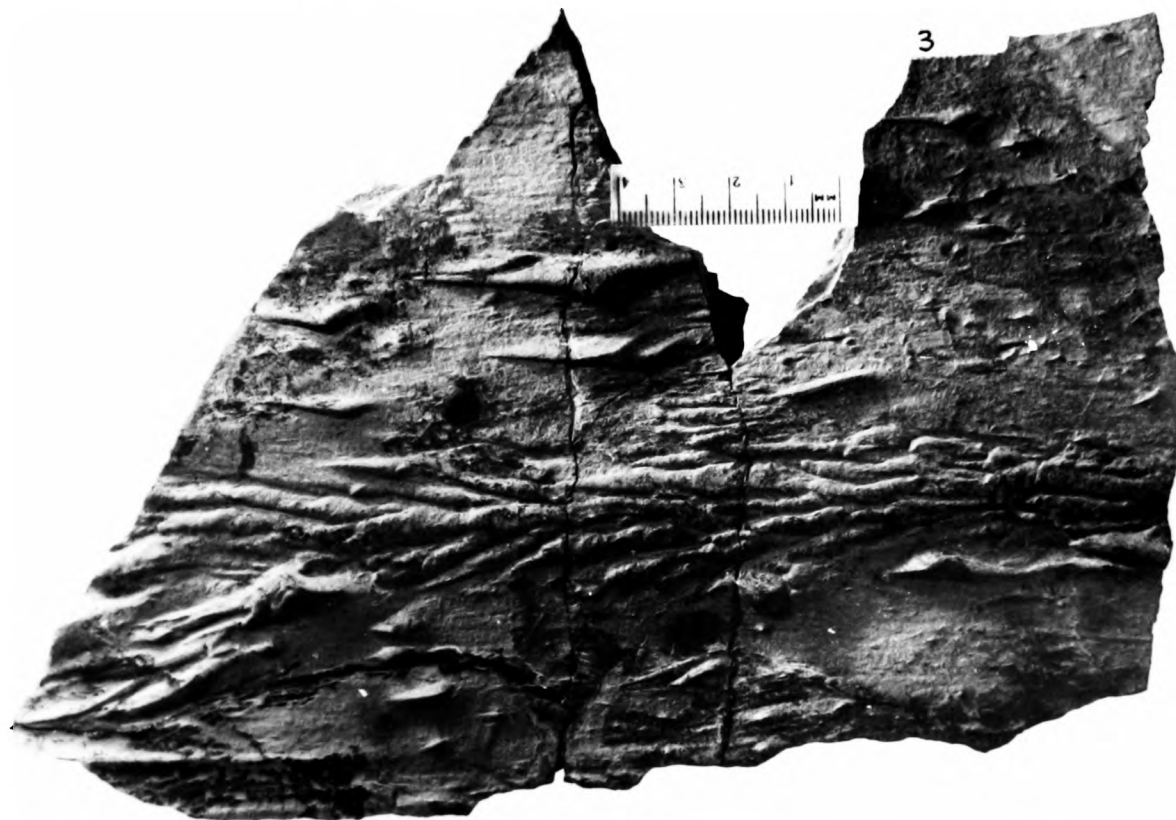


Fig. 2.1Dd

Fleur-de-lys scour structure associated with prod moulds. Unit C. Loc.054.

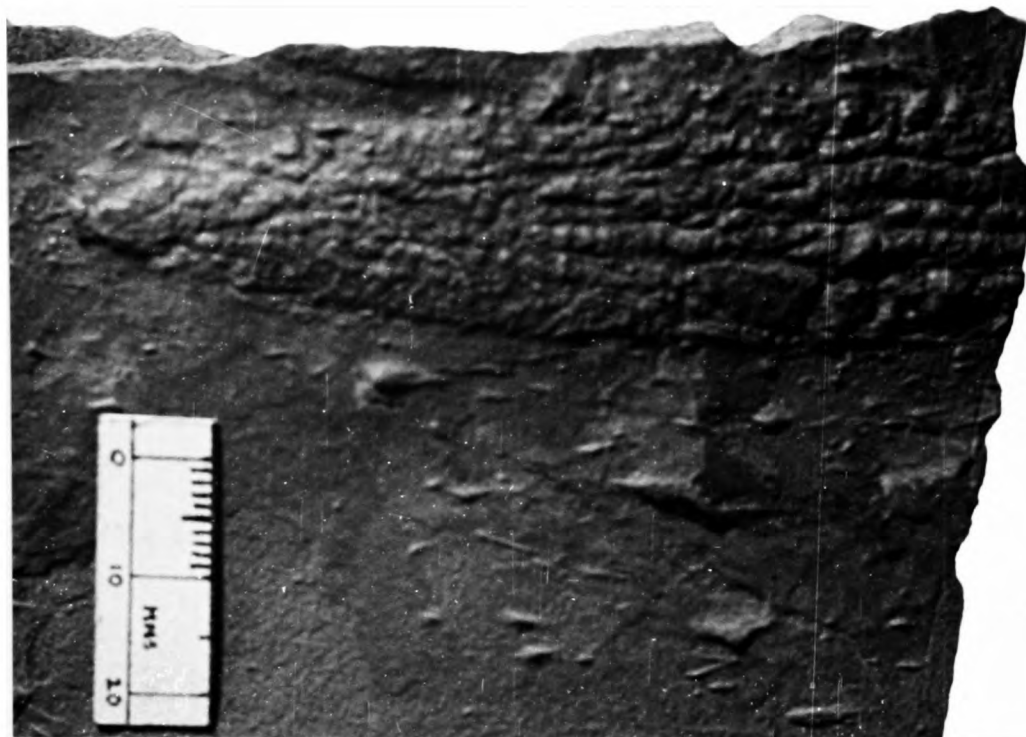


Fig. 2.1Dda

Scour patch allied to longitudinal ridge structure on base of bed associated with small prod moulds. Unit C. Loc.054.

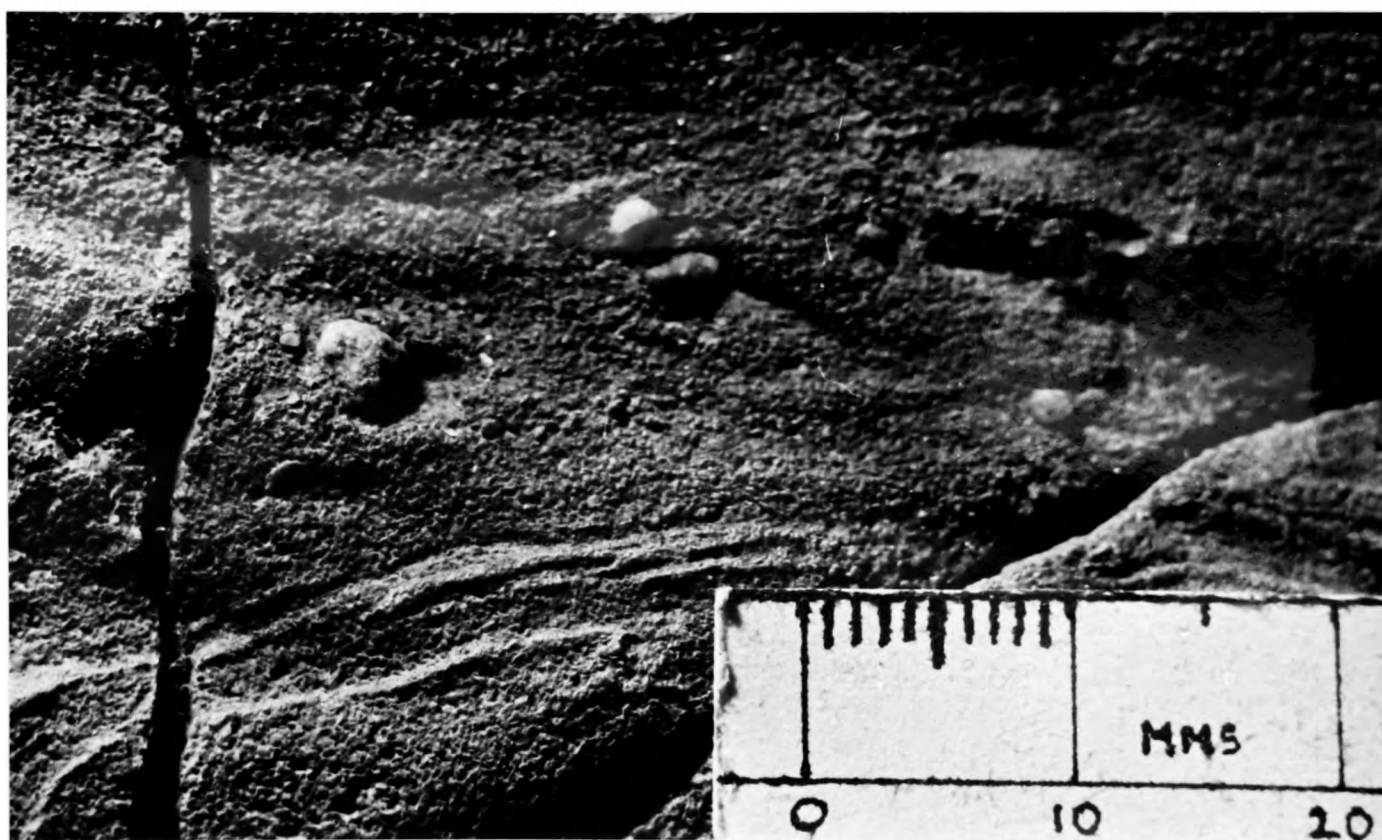


Fig. 2.1Ddb

Scour behind quartz pebbles on the base of a proximal protoquartzite. Unit B. Loc.073.



Fig. 2.1Ddc

Large irregular scours behind the former sites of shale pebbles (smooth areas). Base of proximal protoquartzite. Unit B. Loc.073.

TOOL MARKS.

Tool marks are produced by objects carried by the current impinging on the substrate, and leaving a depression or mark. The depression so formed may subsequently be filled with material deposited from the current and be preserved as a mould on the base of the deposited bed.

Tool marks vary greatly in size and morphology depending on factors such as the shape of the tool, the strength and nature of the current, and the manner in which the tool strikes the substrate. Tool marks can be made by dragging, rolling or saltation of a tool.

Dzulynski and Sanders (1962) classified tool marks into continuous and discontinuous types. In practice all tool marks are discontinuous, but often they persist for several metres without significant change and then they may be considered as continuous. In the present area no exposures permitted examination of the under surfaces of sandstone beds for more than about a metre. If the mark maintained itself uniformly over the surface in question it was considered to be a continuous mark.

CONTINUOUS TOOL MARKS.

Groove marks.

"Groove marks are long, remarkably straight gutter like troughs which run across the outcrop usually with no appreciable change in thickness" (Dzulynski and Walton

1965). The convex feature on the base of the overlying bed should be known as the groove mould.

Groove marks were first described in 1843 from New York by James Hall. They were termed 'Drag Marks' by Kuenen (1951) and the same author described what could be very large grooves as 'Gouge Channels'. These marks were up to 50cm. deep, and .5-2.0 metres across, and had flutes and smaller grooves on their surfaces. Normally groove marks are less than 1cm. and often only a millimetre deep.

The actual shape of the groove produced depends on the nature of the cutting tool and also on the consistency of the substrate. Many objects have been recorded as cutting tools. The most frequent records are of hard or or soft shale fragments (Allen 1960), but there are records of Orthocone Nautiloids (Craig and Walton 1962), Fish bones (Dzulynski and Slaczka 1958) and shell fragments. Allen (1960) thought that plant fragments could also have acted as tools, and in the present study fairly convincing examples of marks produced by stem fragments were found.

Some groove marks give the impression of having been produced by the dragging of a plastic tool over the bed surface. The most likely object would have been a disintegrating shale fragment. Dzulynski and Sanders (1962) figured a specimen with the shale pebble preserved. Similar structures were seen associated with the proximal

protoquartzites at Sprink.

The continuous groove marks or their moulds give the line of flow of the current but not the direction of flow. Groove moulds were the most frequent oriented current structures found in the area. They showed all the typical varieties from the finest striations, possibly produced by single sand grains, up to the structures referred to as gouge channels by Kuenen (1957); Bassett and Walton (1960). Examples of groove moulds are shown in Fig.2.1De.

DISCONTINUOUS TOOL MARKS.

Bounce marks (Wood and Smith 1959).

These marks are best described as doubly terminated groove marks. The tool impinges on the substrate at a low angle of incidence and is dragged along before being gradually lifted back into the current. The bounce mark is therefore "a symmetrical depression tapering and flattening off in both upstream and downstream directions" (Dzulynski and Walton 1965).

There is a gradation from bounce marks to groove marks dependent on the length of the mark, and any division is quite arbitrary. Bounce marks also indicate only the line of the current. Bounce moulds are frequently found on the bases of protoquartzite beds in the area and are probably the commonest bottom structures in the area.

There is a continuous variation in size from moulds less than a centimetre long up to forms gradational to groove marks. Typical forms are seen in Fig.2.1Df.

Prod marks (Dzulynski and Slaczka 1958)(Dzulynski et al.1959)

"Prod marks are asymmetrical, elongated semi-conical or triangular depressions impressed most deeply in a downstream direction" (Dzulynski and Walton 1965). The downstream end may be symmetrical due to the clean lifting of the prodding tool, or it may be twisted to one side due to rotation of the tool in the current as it was wrenched out of the mud. The marks vary in size according to the nature of the tool, the angle at which it strikes the substrate and the force with which it prods the substrate. The marks usually vary from small triangular depressions a millimetre or two in length, up to about 10cm. with a relief of 1cm. (Dzulynski and Walton 1965).

Prod marks are produced by a high angle of incidence with the axis of the tool inclined downcurrent so that it digs into the substrate rather than dragging along it. Craig and Walton (1962) record prods produced by *Orthocone* nautiloids, Dzulynski (1963a) records marks made by fish vertebrae. Many authors have considered shale fragments to be the main prodding tools. Allen (1960), writing on the Mam Tor turbidites, considered both shale and plant fragments to be possible tools.

Prod marks are of frequent occurrence in the present area, and most of the current directions measured were from prod marks. Prod marks are most common on the bases of distal beds that are underlain by shale-mudstone. They have only been seen in thick proximal developments where a substrate of shale-mudstone was present. Thus it seems that the nature of the substrate is critical in the preservation of prod marks. The marks vary greatly but correspond to the size ranges mentioned in the literature. Those measured ranged from 0.15 to 8.0cm. long and up to 1.0cm. deep with symmetrical or strongly twisted ends. In practice it was usually the prod moulds that were studied on the bases of sandstone beds, but careful lifting of a sandstone bed revealed the original marks on the shale surface. Typical prod moulds are shown in Figs. 2.1De-h.

Some of the largest prod marks in the area were found in a section at Shirkley (in the general proximal area) which contained only thin sandstone beds. The prods were associated with scour markings of the longitudinal ridge and fleur-de-lys type (Fig.2.1Dd) and generally turbulent conditions are suggested. The deposited sand beds are frequently less than 10cm. thick and may be represented only by isolated ripples with the bottom structures preserved on their bases. It is suggested that strong turbulent currents formed the prod and scour structures

but that the bulk of the material carried by the current was transported further into the basin before being deposited. The mudstone between the sand beds is green in colour and frequently silty, and no typical black shale-mudstone is seen.

Within the main basin area in more distal sequences of the protoquartzites smaller prod marks are more usual and these are often associated with groove and bounce moulds. Continuous grooves are absent in the thin bedded proximal development.

POSSIBLE TOOLING AGENTS IN THE PROTOQUARTZITES.

In the north Staffordshire basin features were found associated with the sandstones similar to those recorded from the Namurian (R1 Stage) of Mam Tor, Derbyshire by Allen (1960). Shale fragments are found within beds on occasions and plant material is also frequent. Some specimens of prod moulds, particularly those from Loc. 063 Dalehouse Wood show impressions closely resembling the stems of a Calamites like plant. Actual stems (Fig.2.1D1) have been found throughout the area, though they are most frequent and better preserved in the more proximal regions of the sandstones (e.g. Sprink Loc. 072). However, plant fragments were certainly carried by the currents for as far as the sandstones can be traced into

the basin. Thus it seems that the plant fragments could well have acted as tooling agents.

Shale fragments in the form of flat discoidal pebbles are found in some of the sandstone beds. In proximal regions (e.g. Sprink) shale pebbles can be found included in the bases of some sandstone beds (Fig.2.1Dj), and it seems inevitable that these fragments acted as tools. Shale fragments are also found as far into the basin as Pyeclough, but here they are confined to the thicker beds and usually occur within the bed rather than on its base. Maybe these fragments were picked up by the current during its progress rather than being derived from near its source. Shale fragments derived from the source regions would probably have disintegrated before reaching Pyeclough so accounting for their absence in the majority of beds in this region. It is also possible that they were present in the initial stages of flow but disintegrated during or after deposition and were not preserved. The fact that they are found trapped in some beds does at least prove their presence within the currents that deposited the sands.

The more proximal regions of sandstone deposits show occasional beds with included quartz pebbles up to 1cm. in diameter. It is possible that these could have acted as tooling agents, but it seems that they were deposited very quickly as they are found usually in

association with thick sandstones without any shale-mudstone tops but with occasional green or purple clays which contain pebbles (see section 2.2). These deposits were seen at Shirkley, Sprink and Dalehouse Wood. When the pebbles occur on the base of a sandstone they may give rise to an oriented structure, but this is due to scour and not tooling. The lack of quartz pebbles in the distal beds indicates that even if they acted as tooling agents in the proximal beds they were not the only objects acting as tools.

No other objects have been found within the sandstones that could possibly have acted as tooling agents. The only fossils other than plant fragments are trace fossils such as burrows and trails. The burrows were frequently unroofed by the currents, but it is unlikely that the animals could have acted as tooling agents even if they did get caught in the current. Thus it can be argued that the objects responsible for tooling were shale pebbles and plant fragments. It is probable that all the tool marks found associated with the sandstones in the area were produced by one of these agencies.

EVIDENCE ON NATURE OF SUBSTRATE.

No marks were seen that could be classed as brush marks, nor were any chevron marks (Dunbar and Rogers 1957)

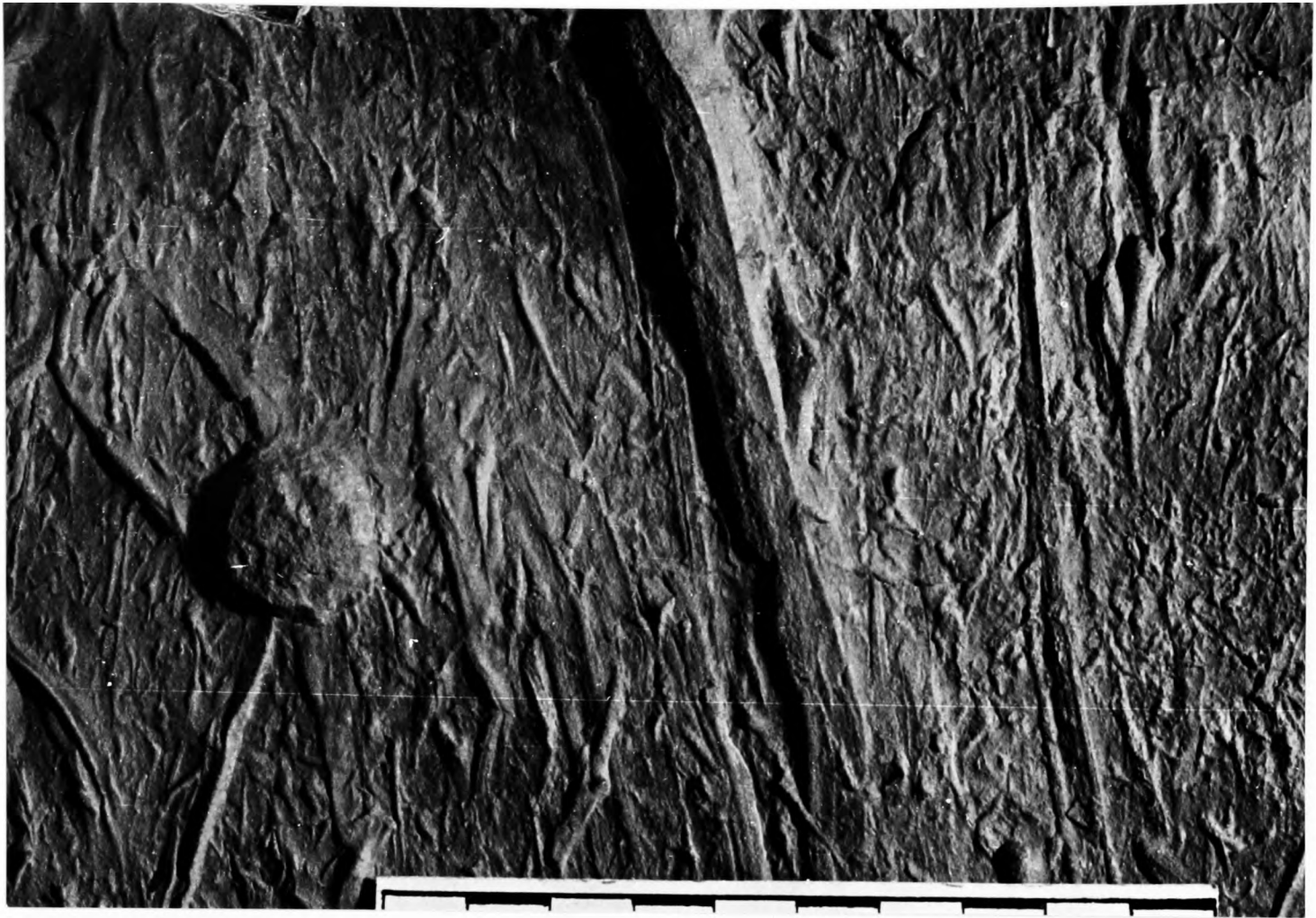


Fig. 2.1De

Groove and bounce moulds together with a sand protruberence
which is possibly the filling of a vertical burrow.

Unit C. Loc.422.

Fig. 2.1Df

Bounce and prod moulds, some with striations suggestive of plant fragments of Calamites type. Loc.063.

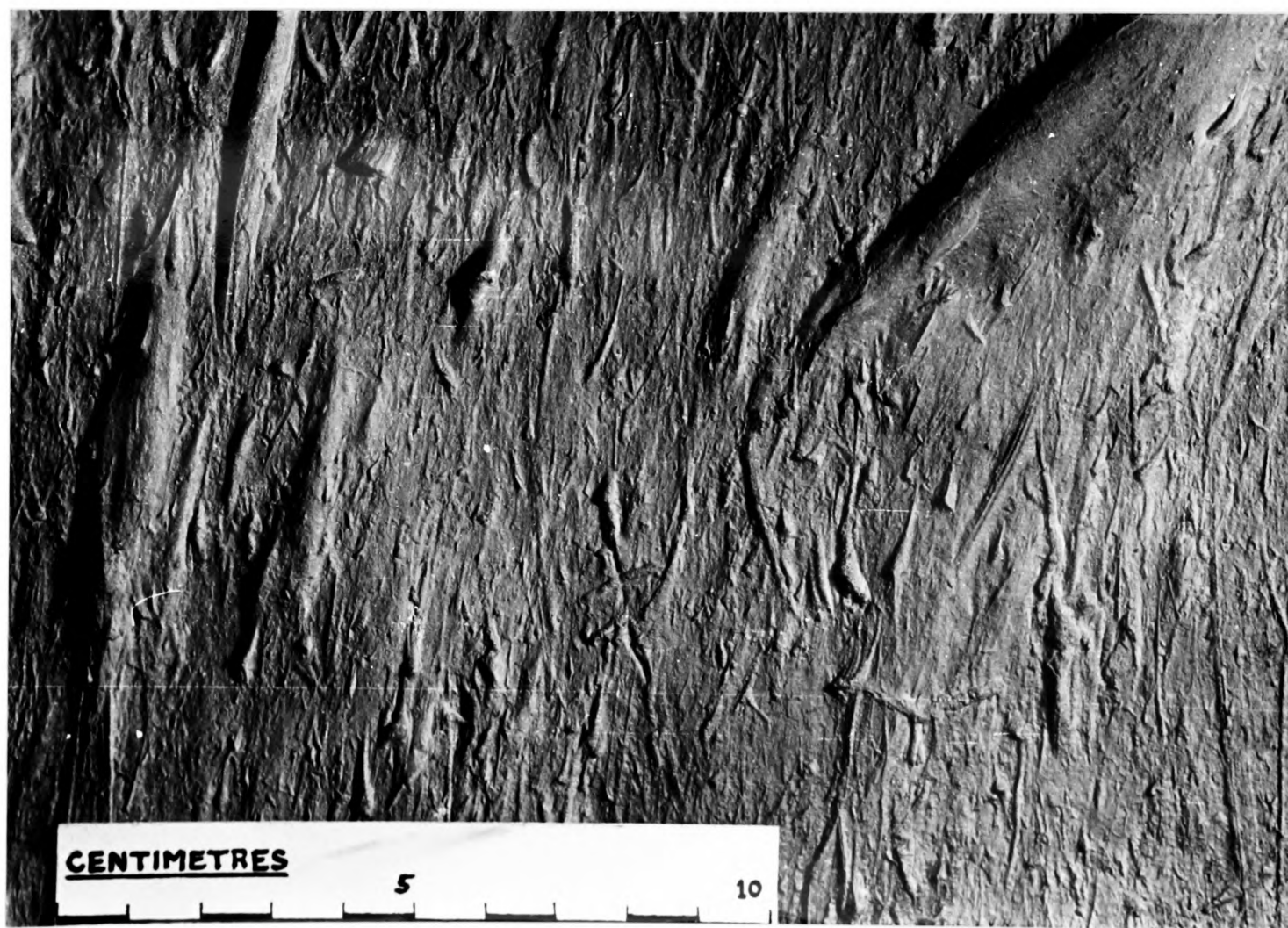




Fig. 2.1Dg.

Repetitive prod moulds possibly made by a plant stem similar to those shown in Fig.2.1Di. Loc.063.

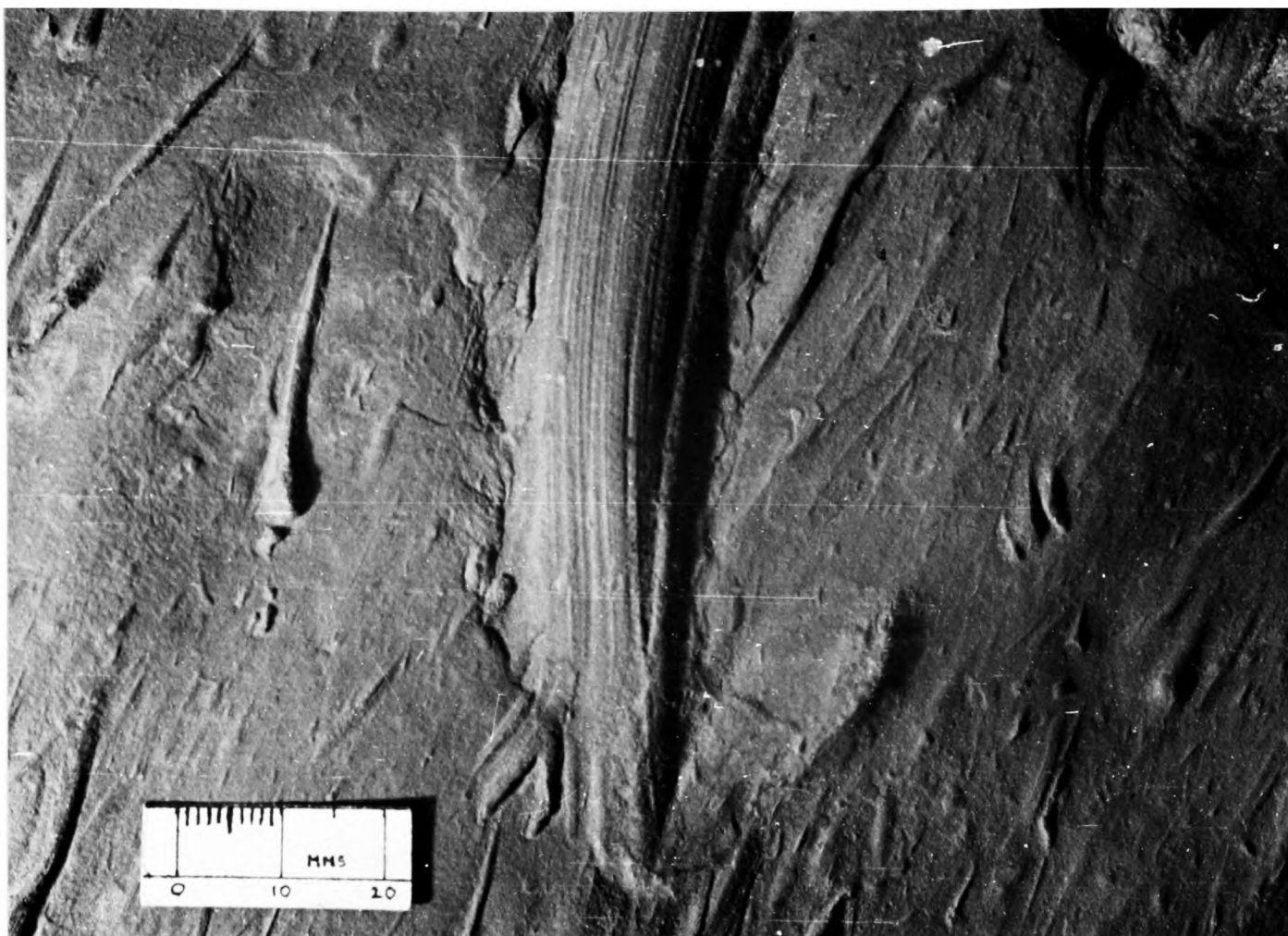
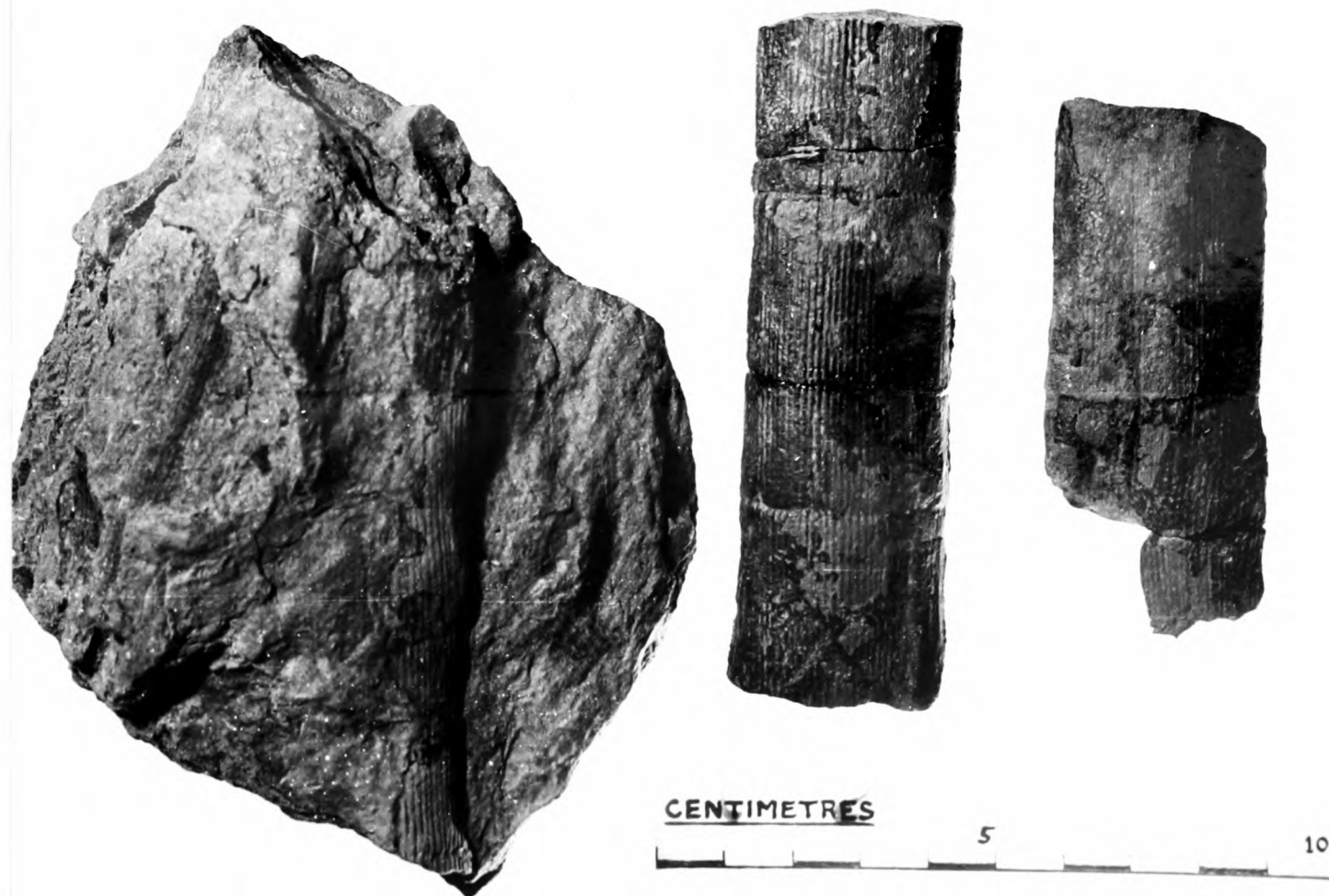


Fig. 2.1Dh

Prod moulds from 2 to 30mm. long and termination of a striated bounce mould. Unit A. Loc.017.

Fig. 2.1Di

Stems of Calamites from between proximal protoquartzites. Such stems probably produced many of the tool markings seen in the protoquartzites. Unit B. Loc.072.



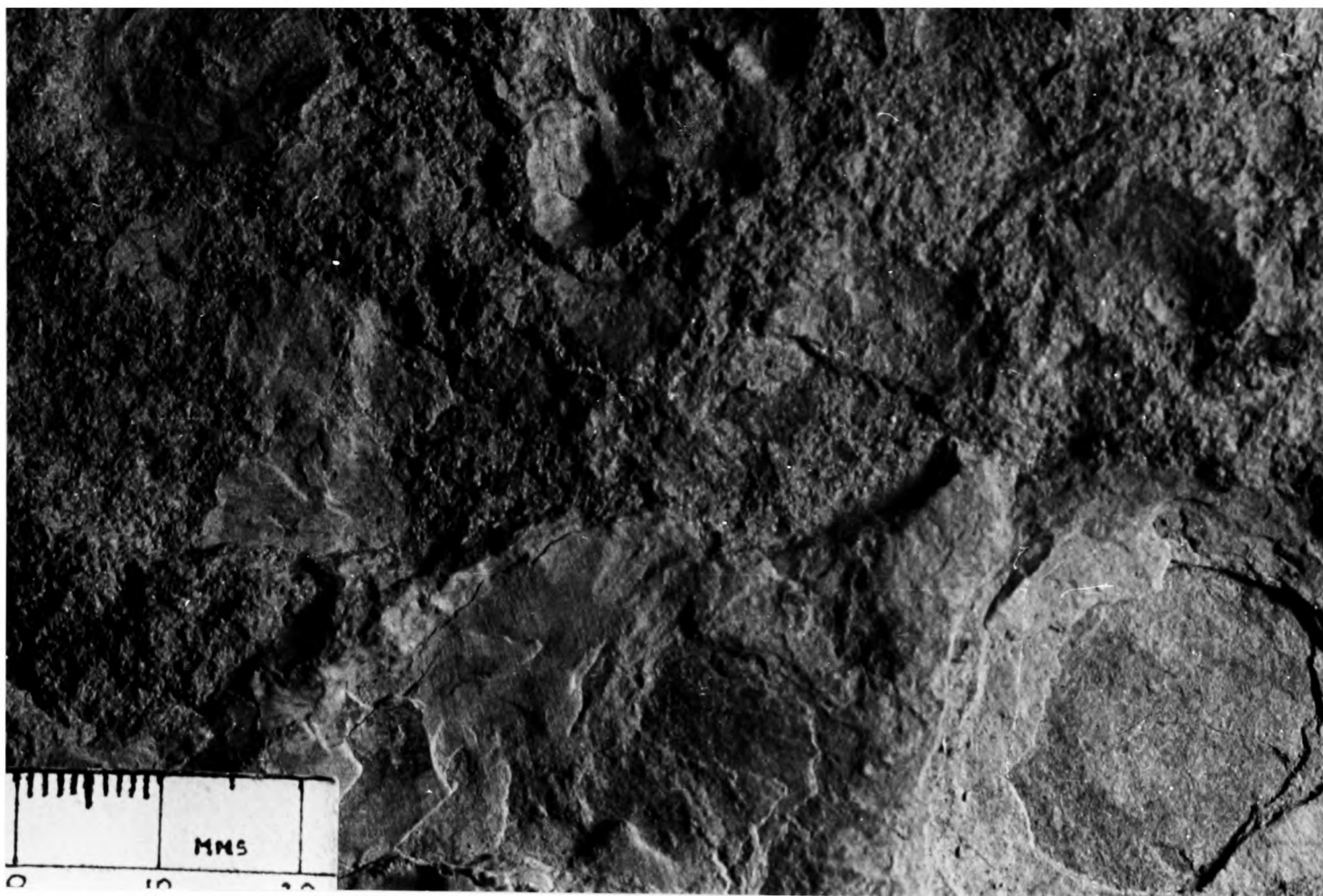


Fig. 2.1Dj

Shale pebbles seen in plan view on a lamination parting of a protoquartzite bed. Such shale pebbles probably produced tool markings on the bases of many turbidite beds. Loc.063.

seen associated with the groove marks. There was also an absence of any transverse tensional cracks or ridges on the substrate. As all these features are associated with some plastic deformation of the substrate, and as no such features were found it is thought that the substrate must have lacked a surface suitable to retain the effects of plastic deformation. The surface obviously had to have a certain amount of cohesive strength to retain the grooves and prods but not sufficient to allow the above mentioned features to form.

LOAD, FLOW AND INJECTION STRUCTURES.

This group of structures is characterised by deep, rounded projections of sand into mud or laminated sediment. Many names have been given to such structures often according to the mode of formation envisaged by the author. Shrock (1948) described structures under the name 'Flow cast' as he considered horizontal flowage of material to have taken place. The term load cast was used by Kuenen (1953) as the form of structures suggests that some vertical sinking has taken place. It is however quite obvious from a study of the literature that there are forms produced by horizontal flowage combined with the action of gravity loading, and forms produced purely by gravity loading. This was realised by Greensmith (1956)

and by Kuenen and Prentice (1957) who considered flow structures to form from " the horizontal movement of the base of a bed of sand during or after its emplacement on a mobile substratum, combined with sinking and ploughing up of the latter", and load casts to be " due to the vertical adjustment of the basal material to unequal loading".

This would seem to be the only part of the story as there is also a gradation from types where only the base of the bed is involved to those where the whole of the bed is disrupted into slump balls (Allen 1960) or flow rolls (Prentice 1956, 1960) which have a special orientation with regard to the current. The bed may disrupt into pseudonodules (Kuenen 1958) which are non-oriented with respect to any current. Thus many of the load and flow structures described in the literature are genetically related.

Classifications of load, flow and injection structures.

Dzulynski and Walton (1965) classify load structures "according to the processes leading to unequal loading."

These are: 1. Filling of current marks.

2. Differential deposition, as in ripple marks.

3. Formation of a wavy interface by the passage of shock waves.

The first two of these groups are fairly easily distinguished except when the loading process has gone

so far as to obliterate the original marks that initiated the load structure. The third is virtually impossible to prove, and even if the structures can be produced by shock in the laboratory (Kuenen 1958) it does not prove that geological counterparts were produced in this way. For these reasons it is considered that the above classification scheme is unsatisfactory. A better scheme should take into account the physical variables of the environment that lead to the formation of the different forms of load and flow cast.

Kuenen and Prentice (1957) in their classification of sedimentary structures separate load and flow casts into two groups, but also note that combinations of the two can occur. They also introduce a group due to 'differential compaction' - which can distort laminae around a flute and so give it a resemblance to a loaded flute or an ordinary load cast, especially when seen in section. Dzulynski and Walton (1965) also separate load and flow structures but have little to say about the latter - obviously they must be scarce in the Carpathian Flysch.

Flow structures.

Flow structures have been described many times in the literature and only a brief survey of the types mentioned and a comparison with those in the present area is given.

Flow structures vary in size from small overfolds and balled up structures affecting only the bottom 3cm. of an 8cm. bed (Prentice 1956) or equally small structures forming a discontinuous bed, (Shrock 1948 Fig.116c) up to large structures such as the flow rolls of Sorauf (1965). These are folds protruding from the base of a sandstone which are oriented perpendicular to the current movement and are from 4 inches to 3 feet long. These are similar to structures described by Kuenen and Prentice (1957). The large scale equivalent of the discontinuous bed of Shrock (1948) is the structure described by Allen (1960) as 'Slump Balls' where recumbent folds with axes perpendicular to the current direction are piled one on another to form balls up to 50cm. in diameter. The individual folds may have distorted groove moulds on their surfaces perpendicular to fold axes showing that the current did some cutting of the sediment prior to deposition and that during deposition and still under the influence of the current the bed and underlying mud deformed plastically. It can also be argued that the later deformation is due to the palaeoslope and not the current direction. If the bed has a planar top showing that flow movements had stopped before the deposition of the final part of the bed then it is probable that the flow structures were produced under the influence of the current. If the top of the bed is also distorted then it is possible that the rolls

were produced by a down slope movement after deposition, i.e. sliding or slumping. Thus care must be taken to distinguish the two as one is connected with palaeoslope direction and the other with current direction.

Load structures.

Load structures are very variable in size, and mode of formation. They can arise by the distortion of previously existing structures such as groove marks or flute marks (Kelling and Walton 1957). These structures marked the sites of slight overloading of the substrate and so were the points of initiation of load structures. Unequal loading can result from ripple marks which may become loaded in remarkable ways (Dzulynski and Walton 1965). Shock waves producing sudden liquefaction of sands may also be a cause of loading. This must obviously occur post-depositionally and possibly at depth in the sediment.

A syndepositional load cast may sometimes be distinguished by its coarse fill but care must be exercised not to confuse it with a scour mark of a non oriented variety. Extreme cases of load structures are pseudo-nodules in which the whole of the loading bed breaks up and sinks into the underlying bed as pseudo-nodules of sand (Kuenen 1958; Dzulynski and Walton 1965, p.161).

Load cast morphology has been described in detail by Sullwold (1959, 1960) but as Holland (1960) pointed out some of his terms are unnecessary as terms are already

in existence for features he describes. His terms load pocket and load fold find favour with Dzulynski and Walton (1965) who pointed out that the structure was not formed as a mould or cast as in the filling of a current mark and proposed replacing 'cast' by 'structure', and an amended diagram is suggested below (Fig.2.1Dk). Some load structures have their associated flames oriented in a common direction. This suggests that there was still some horizontal movement taking place in the overlying bed, and this represents a transition towards flow structures. It is probable that the two form a continuous series.

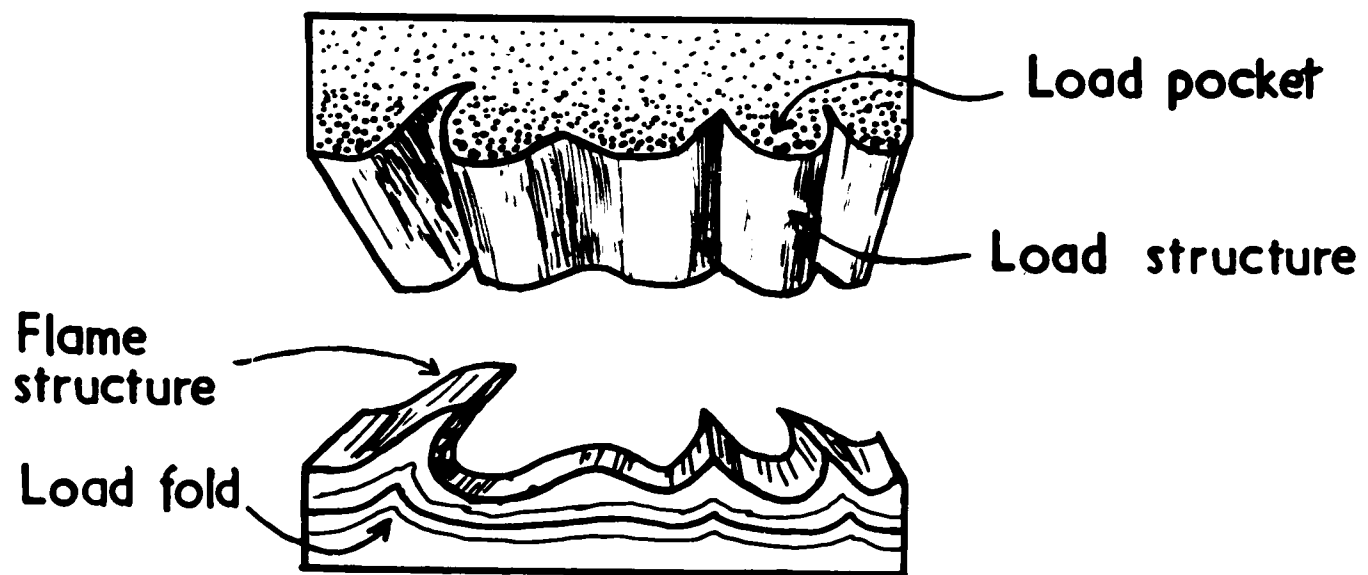
Load and flow structures on beds with planar tops.

The factors that affect these structures are:-

1. Gravity.
2. Thickness of loading bed.
3. Physical state of substrate on which bed is deposited.
4. Lateral movement.
5. Depositional slope.

There are thus two continuous variables to be considered. One being from pure vertical loading on no slope to a strong lateral movement of material on a steep slope. The other most important variable is the state of the surface on which deposition takes place. This can vary from a hard planar surface capable of withstanding scour by the current and not yielding to loading after or during deposition, to a surface which

Fig 2.IDk



Load morphology diagram modified from Sullwold 1959.

yields immediately to loading to such an extent that the whole bed is broken into flow rolls or pseudo-nodules syn- or post-depositionally. The thickness of the bed is important in that a thicker bed can apply a greater load so increasing the likelihood of load deformation taking place. The following tables give a general idea of the structures produced by combinations of these variables.

Effects of frictional drag on beds below depositional surface.

1. Hard bed. Clean cutting leading to fracture and sand injection.
2. Soft bed. Destruction or crinkling of lamination, drag folding of lamination. Flame production.
3. Very soft. Liquefaction - incorporation in overlying bed as very elongate flames.

Substrate type	Vertical loading only.	Downslope gravity component or strong horizontal flow.
Hard Planar Bed		Shallow Scours and Tool marks. Predepositional cutting and scouring of sand
Plastic Bed.	Shallow load casting of ripples flutes or grooves	Deeper scour and tool marks plus plastic deformation around them, wrinkles etc.
Soft Bed	Loaded ripples flutes grooves prods etc. (Post depositional loading) Ordinary load casts - non oriented	Load casts with oriented flames Syn depositional loading of ripples flutes grooves prods etc. Flow casts Frondescant marks
Very Soft Bed	Bed disrupts to pseudo-nodules	Detached flow rolls piles of flow folds discontinuous beds
	Post depositional	Pre-Syn depositional

Effects within beds of loading - current drag phenomena.

<u>Loading Postdepositional.</u>	<u>Loading Syndepositional under Horizontal Flow.</u>
No textural difference in loads; non oriented flames; disturbance of lamination.	Load pockets with coarse grains trapped; oriented flames; lamination undisturbed in upper part of bed.
Maybe some pure load convolutions - no angular discordances of material in synclines.	Convolute lamination with synclines filled by material deposited with angular discordances, cross-bedding in synclines, erosion from anticlines, dying out downwards and maybe upwards, associated with ripples.

LOAD AND FLOW STRUCTURES ASSOCIATED WITH THE PROTOQUARTZITES.

Load structures.

Non-directional load structures are most frequently seen on the bases of sandstone beds that are closely underlain by another sandstone or separated from an underlying sandstone bed by only a thin shale or mudstone bed. Frequently two sand beds may be welded together and only small remnants of the original loaded contact remain. Frequently it is impossible to tell whether the loading took place syn- or postdepositionally. Some beds, with non-oriented bulbous load casts in which the load pocket contains grains similar to those in the rest of the bed, and in which there is no clear lamination of the sand, appear to have been loaded postdepositionally.

In other cases the load pockets are filled with much coarser material than that which comprises the rest of the bed, and the bed may be well laminated near the base. In these cases loading has taken place syndepositionally and there is frequently evidence of syndepositional movement, such as crinkling of laminations underlying the bed as seen in Fig.2.1D1. In this specimen the coarser grains are clearly concentrated in the load pockets, and the load structures are clearly affected by horizontal flowage of the sand bed. The upper part of this bed is perfectly evenly laminated and it seems probable that the

loading and distortion of the basal part of the bed had taken place before the evenly laminated part of the bed was deposited, in other words the loading was syndepositional and took place under the influence of both horizontal and vertical movements. Another example of a load cast filled with coarse material is shown in Fig.2.1Dm.

Frondescent structures.

These structures are very variable in form. Basically they consist of protrusions from the base of a bed usually extending below the level of flute or groove moulds. They are generally flat in section with a convex edge in the direction of current flow. The front edge is crenulated and the surface generally bears radiating ridges which bifurcate outwards to the margin. In plan they vary from elongate in the current direction to short and broad and rarely they approach a circular form with radial ridges. The structures vary in size from a few centimetres up to over a metre across, but their relief from the base of the bed is not usually very great.

The structures have received many names in the past all of which give an idea of their morphology, 'frondescent marks' Ten Haaf (1959); 'Cabbage leaf structures' Kuenen (1957); 'Feather-like marks' Ksiazkiewicz (1958); 'deltoidal marks' Birkenmajer (1958), are some of the synonyms of this structure.

The marks seem to be formed by the syndepositional sinking of the depositing sediment through the surface of the substrate only to spread out within the substrate at a shallow depth - maybe on top of a slightly harder bed or at a depth where the mud was sufficiently hard to resist intrusion by the depositing material. The ridges on the surface represent the diverging lines of flow of material from the source to the margin of the structure.

This has been observed experimentally with a gelatine substrate and seems to be the only reasonable explanation for their formation (Dzulynski and Walton 1965, p.222).

Assuming that they form by sand intrusion into mud the term structure should be used as at no time is a predepositional mark formed. This fact was referred to by Dzulynski and Walton (1965 p.132). Their frequent orientation suggests that they are formed syndepositionally but comparisons between orientations of frondescent marks and other current directional structures may reveal some to be of postdepositional origin connected with loading on a slope. Some frondescent structures were observed in the present area but usually they could not be fully exposed so their directional nature could not be ascertained. They are typical forms of the structure.

Flow rolls.

Flow rolls are not common in the protoquartzites but a few examples have been found. The simplest form

is illustrated in Fig.2.1Dn and shows the base of a laminated bed deformed into a fold and the top part of the bed adjusted by erosion and deposition to give a planar top. The best examples of flow rolls were seen at Loc. 015 Hurdlow where a protoquartzite bed of about 50cm. average thickness was found to have well oriented irregular flow rolls on its base. The shale-mudstone into which the rolls are pushed has been squeezed up between the rolls and sometimes incorporated within the sand bed. Fig.2.1Do illustrates the features associated with this bed.

The axis of a flow roll may be perpendicular to the direction of the current that deposited the bed or it could have a relation to the local palaeoslope direction. The assumption that the current flowed directly down the palaeoslope may not be valid, and thus the axial direction of flow rolls may indicate neither the palaeoslope or current direction but a combination of the two. Occasionally groove or prod marks may be found preserved on the bases of flow rolls and in this case the relation of the axial direction of the flow roll to the current direction can be determined, and this may indirectly give some indication as to the palaeoslope direction.

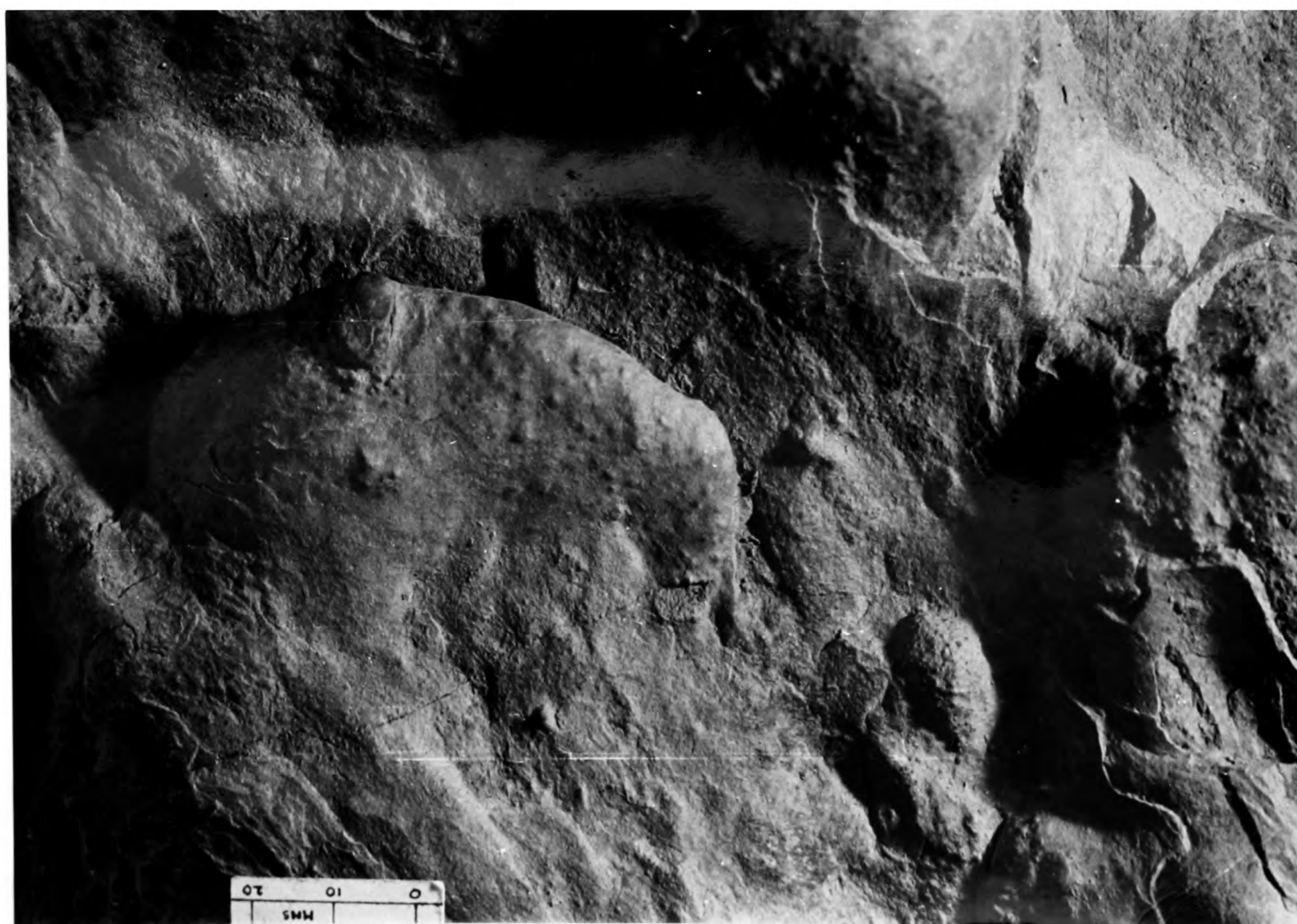
Sand injection structures.

One example of a sand injection structure was seen at Sprink where a sand bed only 4-5cm. thick had stringers

of sand extending from its base down into the underlying shale-mudstone (Fig.2.1Dp). The underlying shale-mudstone contained a few sandy laminations which showed that the substrate had been contorted and cracked in places. The sand from the overlying bed extended down into the cracks and sometimes horizontally parallel to the laminations in the shale-mudstone. The base of the sand bed is very irregular and individual coarse sand grains protrude from it giving it a distinctive 'grain texture' not seen on any other bed (see Fig.2.1Dq). It seems that the substrate was distorted and fractured during the passage of the current and the fractures and irregularities thus produced filled with sand. The planar top to the sand bed also lends support to the syndepositional origin of this structure, but it must be remembered that the top of the bed could have been modified by currents after the distortions took place.

Fig. 2.1D1

Load structures containing coarse grains and pebbles.
Note distortion of previously deposited laminae around
load structure at lower right. Unit C. Loc.422.



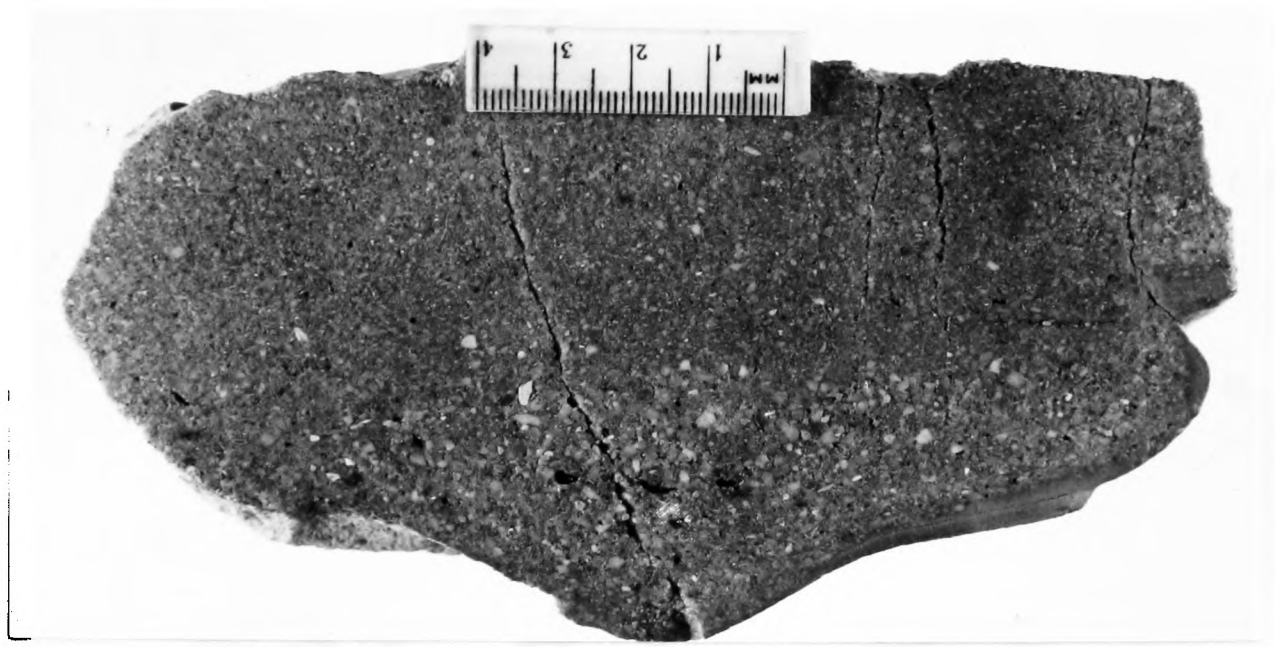


Fig. 2.1Dm

Load structure with coarse grained fill. Unit B. Loc.072.

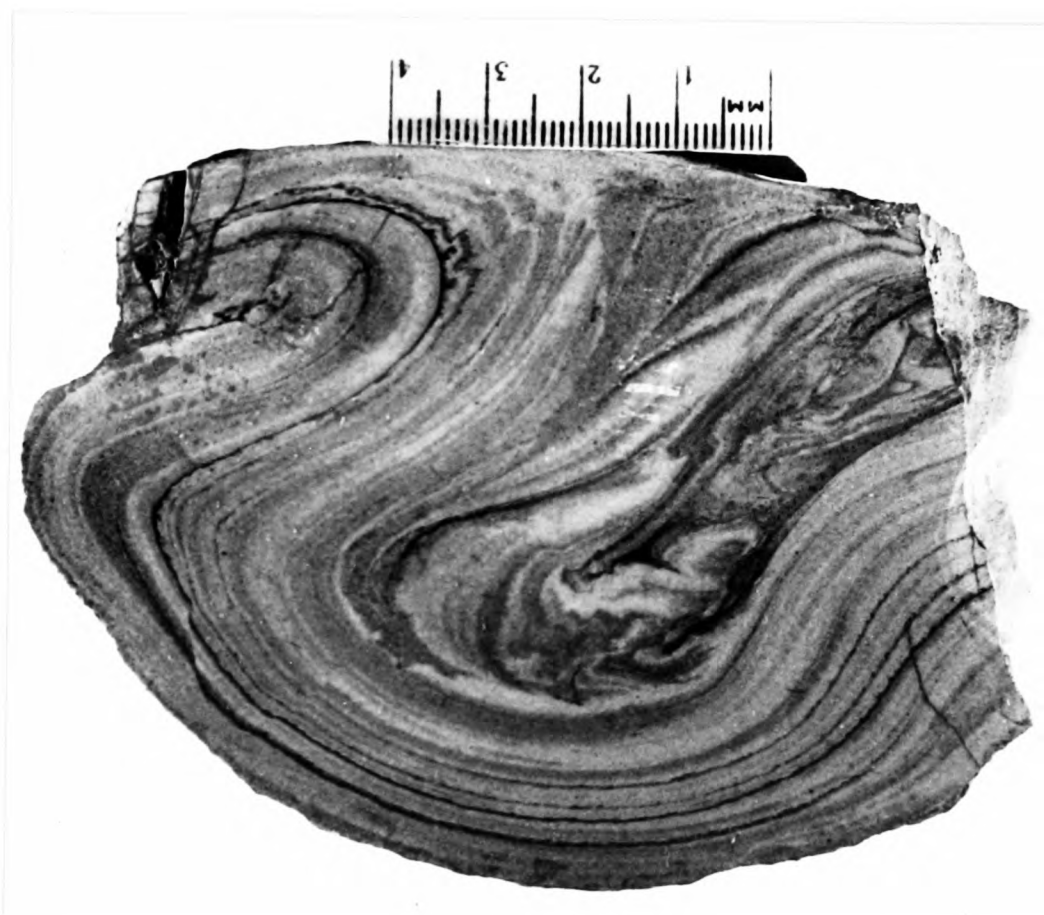


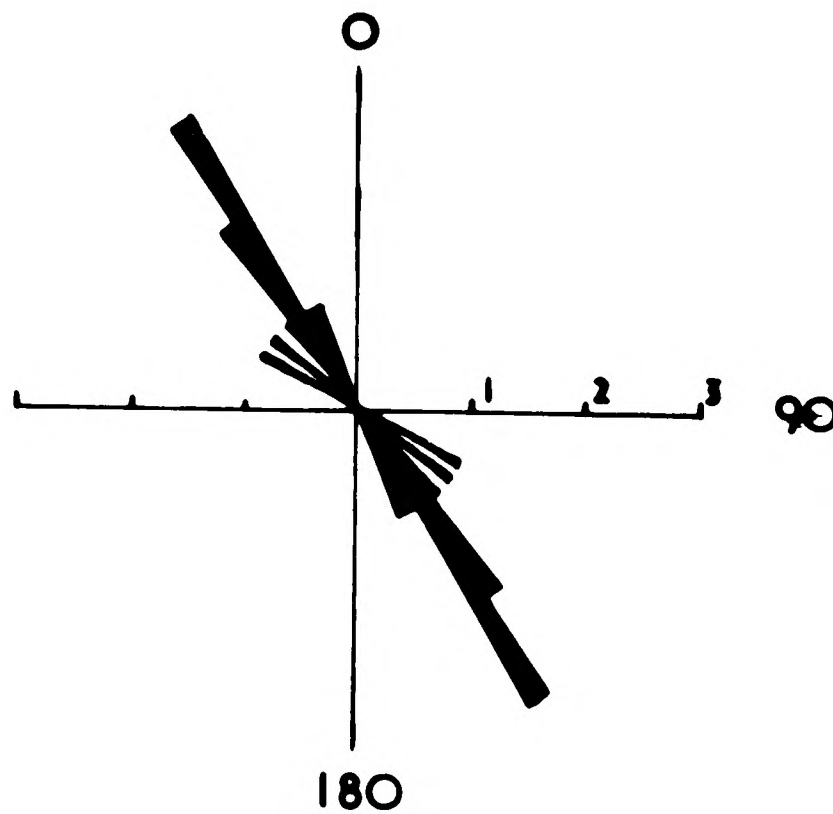
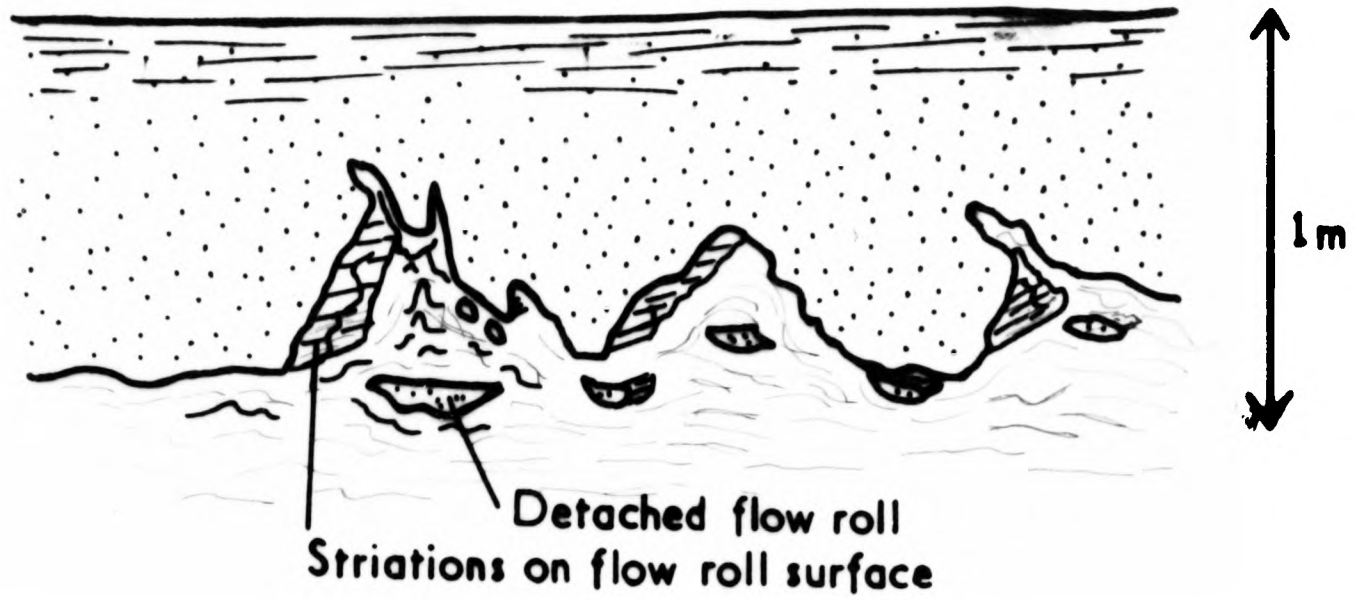
Fig. 2.1Dn

Flow roll on base of bed. Unit B. Loc.464.

Fig 2.1 Do

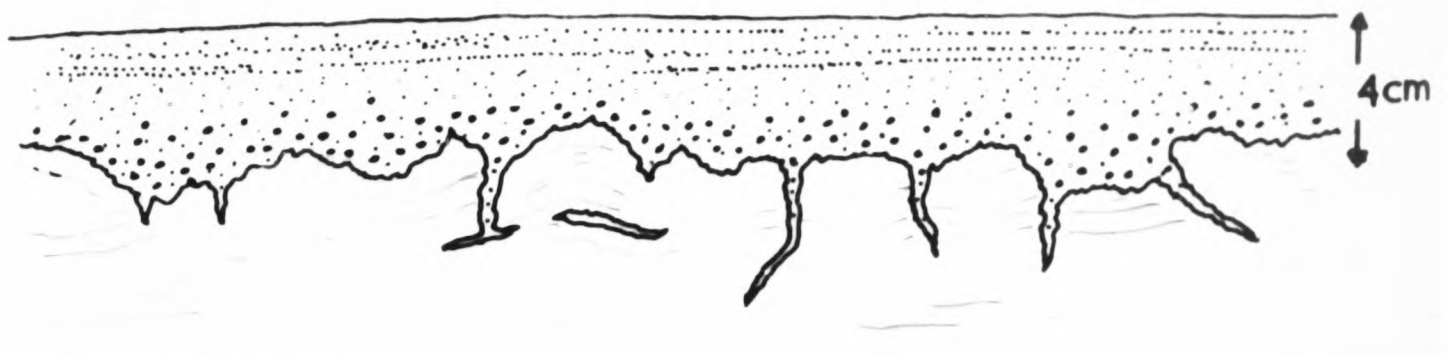
FLOW ROLLS

Loc O15 Hurdlow



Orietaion of flow roll axes and striations

Fig 2.1 Dp
SAND INJECTION STRUCTURE



Loc O72 Sprink.

Stringers of sand extend down into the underlying shale mudstone which has been fractured and slightly distorted.

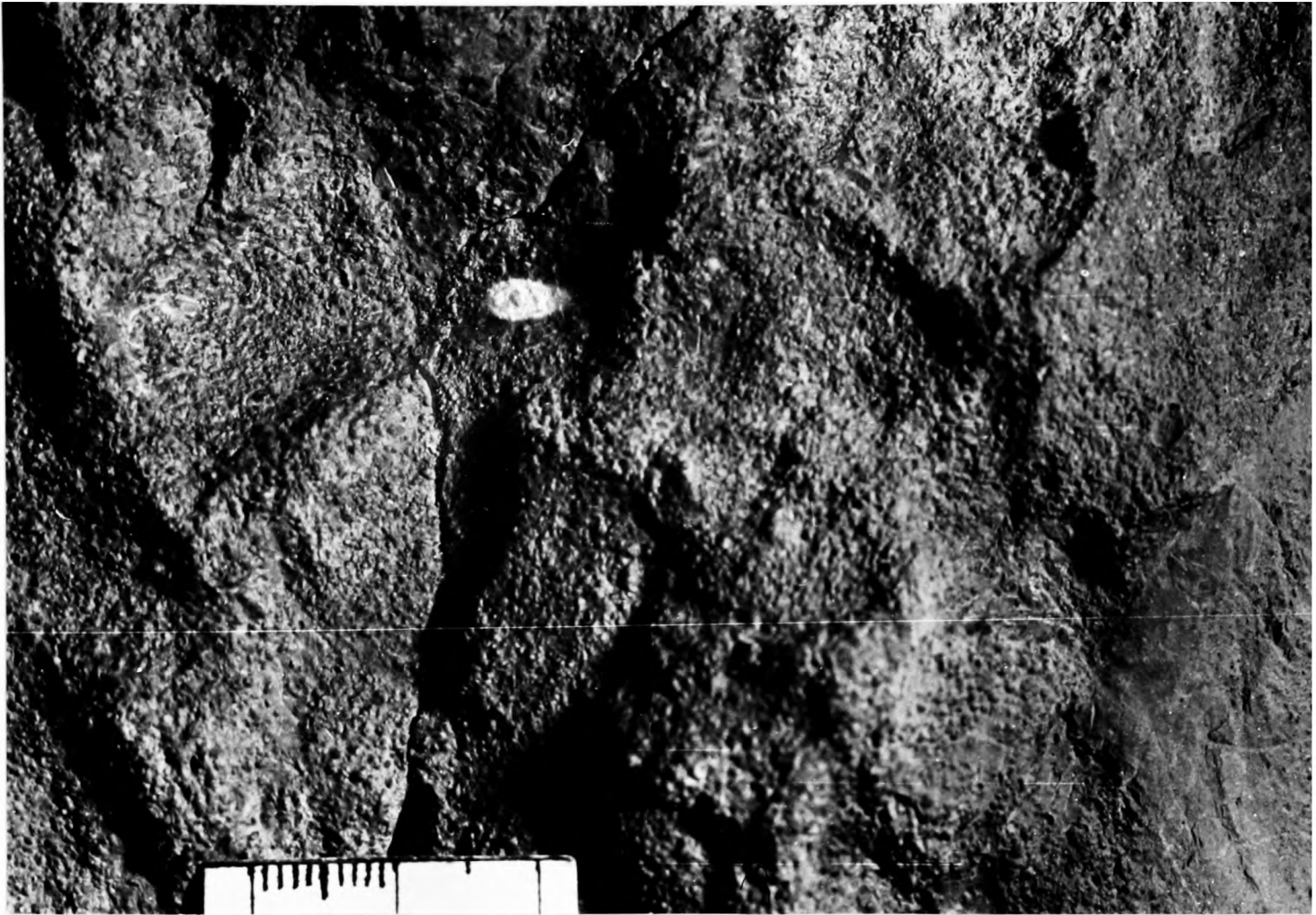


Fig. 2.1Dq

'Grain texture' on irregular base to bed from which downward injection of sand had taken place (see Fig.2.1Dp).

Unit B. Loc.072.

NON-DIRECTIONAL SOLE STRUCTURES.

The soles of the protoquartzitic turbidites frequently show structures that are apparently of organic origin and represent the sand filled burrows of some presumably soft-bodied organisms. Several general names have been given to such structures in the past, e.g. hieroglyphs (Fuchs 1895), trace fossil (Simpson 1957). Many names have been given to different types of trace fossil (see Häntzschel 1962) for descriptive purposes but from the present point of view the mode of origin of these trace fossils is more important.

Holdsworth (Thesis 1963 p.271-281) discusses the trails which he found on the bases of protoquartzite beds in the Longnor-Hollinsclough-Morridge area, and concluded that many of the trails were predepositional in nature and were preserved by the unroofing of the burrow by erosion and subsequent filling by sand from the depositing current. The markedly coarse grained fill observed in some burrows confirms this interpretation (Fig.2.1Dr) (Holdsworth Thesis 1963 p.274 Pl.54,61).

The author is able to confirm this mode of formation for the majority of sole trails seen in the area which are of the type shown in Fig.2.1Ds and in Holdsworth (Thesis 1963 Pl.62, 54) or of the Granularia type Fig.2.1Dt (Ibid. Pl.68-70). In none of these specimens can any

organic disturbance be found in the deposited sand bed and the fill of the trace fossils is frequently coarser than that of the rest of the bed, indicating that the unroofed burrows trapped the coarsest grains before general deposition from the current started. The type of trail described as Granularia by Holdsworth (1963 Pl.68-70) appears to correspond well with that described by Seilacher (1962) but does not closely resemble the forms illustrated as Granularia by Häntzchel (1962 p.w194 Figs.123-1 and 123-5).

One structure observed (Fig.2.1Du) is possibly a trace fossil and consists of small rounded projections from the base of a bed up to 1.5mm. in diameter. The base of the bed on which they occur is obviously erosional, bearing tool markings, and there is no internal disturbance of lamination in the sand bed. The protruberences probably represent small unroofed and sand filled vertical or near vertical burrows. Similar structures, but with a diameter of up to 5mm. were observed on the base of a bed at Loc. 018 Hurdlow and also appeared to be predepositional.

Cases of postdepositional sole trails due to organisms burrowing at or through a sand-mud interface are rare in the present area. A single example was found by Holdsworth (Thesis 1963 Pl.63-65) where pyrite filled burrows occurred on the base of, and within, a sand bed. A further example of a sole marking produced by an organism after deposition

of the sand bed is illustrated by Holdsworth (Thesis 1963 Pl.66) and consists of a conical protruberence on the base of a bed with a relief of 1.1 cm. which shows in section a relation to a break in lamination extending through the laminated and rippled bed. This structure appears to be due to an organism burrowing vertically through a sand bed. One example of a postdepositional burrow entering the top of a sand bed was observed by the author but it did not affect the base of the bed.

Holdsworth (Thesis 1963 p.281) considers that it is impossible to tell whether some of the structures he observed (Granularia and Phycosiphon) were produced pre- or postdepositionally. The present author considers that the majority of the trails are predepositional in origin and due to unroofing of burrows in soft sediment by erosive currents which then cast the burrows in sand during deposition.

Beds showing traces of burrowing are fairly common in the more distal areas studied but become less frequent as the proximal areas are approached, and in the Sprink, Shirkley and Endon sections such structures are very rare. It seems probable that the organisms preferred to burrow in the basinal mud, which was probably rich in organic detritus, and since such mud is more abundant in the distal areas it is logical that burrow structures should be more frequent there. It is also possible that the



Fig. 2.1Dr

Coarse grained fill to unroofed burrow on base of laminated protoquartzite bed. Unit C. Loc.351.

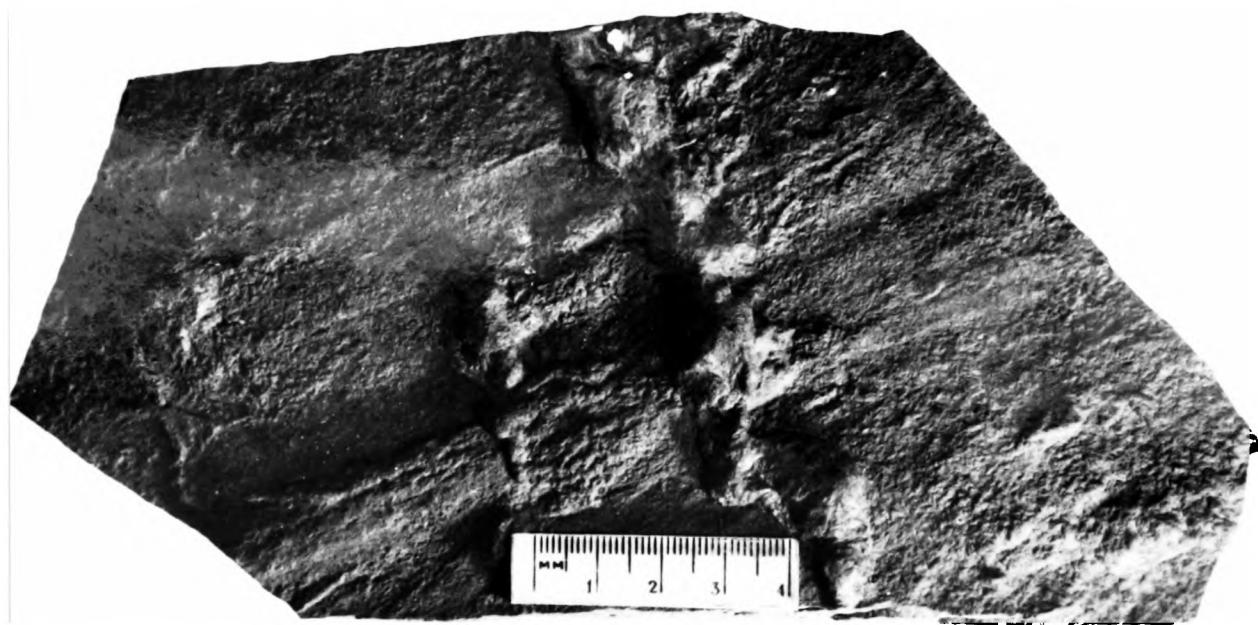


Fig. 2.1Ds

Large irregular unroofed burrow on base of protoquartzite bed. Unit B. Loc.110.



Fig. 2.1Dt

Unroofed burrow of branching 'Granularia' type on base of protoquartzite. Unit B. Loc.110.

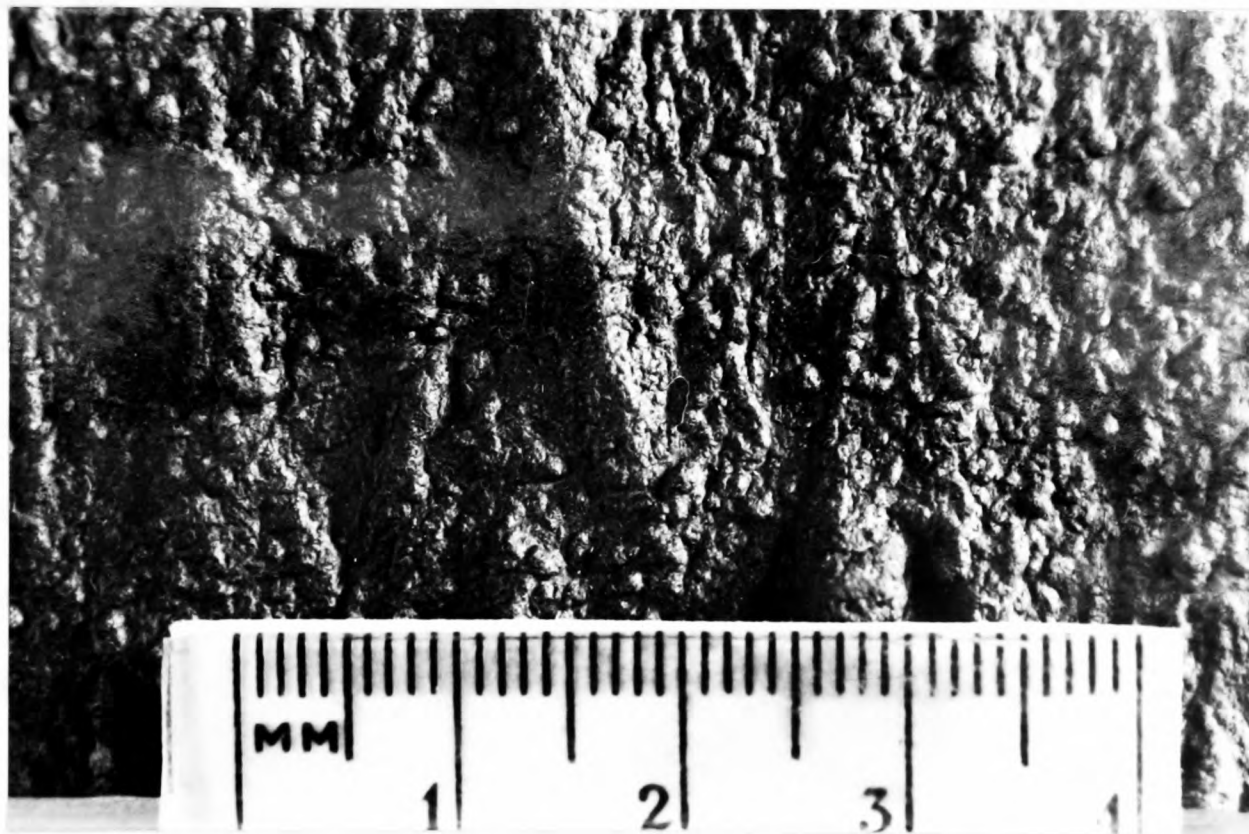


Fig. 2.1Du

Small round projections on the base of a protoquartzite bed which probably represent the fillings of the ends of small partially eroded vertical burrows. Unit B. Loc.110.

currents in the proximal areas eroded a greater thickness of the substrate and hence burrows would be less likely to be preserved.

SLUMP STRUCTURES.

Slumping, the downslope movement of previously deposited beds under the influence of gravity, results in extensive deformation of the slumped sediment. The slumped sediment mass often contains numerous folds produced by rolling and plastic deformation of original beds, as well as pull-apart structures caused by stretching and fracturing of the more competent beds and laminations contained in the slumped mass.

In the area possible deposits were found at Hurdlow, and in the outcrops at Hawkshead Quarries and Loc. 360 Croker. At these last two localities the exact stratigraphical level of the slumped protoquartzites is not known, but at 360 Croker it is probably the B unit of protoquartzites. At Hawkshead Quarries there is no real evidence for the stratigraphic position, but the lithofacies of southerly derived protoquartzites exposed is similar to that found in E2 in other parts of the area. At Loc. 015 Hurdlow about 3 metres of slumped beds are rather poorly exposed. The deposit consists mainly of distorted shale-mudstone with occasional irregular sand bodies which

appear to represent pulled apart fragments of originally continuous beds. One such fragment is shown in Fig.2.1Dv and illustrates microfaulting and partial destruction of the original lamination due to pulling apart of the bed. No directional information could be obtained from this deposit.

At Hawkshead Quarries Loc. 370 about 12 metres of slumped material is exposed overlain by evenly bedded protoquartzites with some beds up to 80cm. thick. The slumped strata consist of shale-mudstones and sandstone beds with shale-mudstone dominant. The sandstone beds are extremely distorted and consist of isolated slump-fold hinges (Fig.2.1Dx), fragments of pulled-apart beds (Fig.2.1Dw), inverted fragments of sand beds etc. The shale-mudstone, which was laminated, has a lineation on bedding surfaces parallel to the axes of the slump folds. The directions of the lineations are very variable but the majority are oriented roughly north-south (Fig. 2.1Dy). The fold axes preserved in the slump are mostly disconnected from the limbs of the folds which occur as isolated pulled-apart bed fragments. Siderite is present in the slumped beds, one pull-apart fragment being preserved in siderite. It is not possible to tell whether the siderite development was pre- or post-slumping. Also within the slump a piece of coalified wood 24cm. long was found in such a position that it must have been rigid

at the time of slumping and converted to coal later.

During slumping the sand beds behaved plastically to a certain extent before breaking into fragments. Thus pulled-apart blocks are slightly thinned at their ends. The attitude of the slump folds indicates slumping from the west.

The exposures in Fox Bank Quarry (Loc. 360 Croker) show very evenly bedded protoquartzites which can be seen to continue unbroken for about 100 metres over the quarry face. The sandstone beds are grouped in units of from 10-25 beds, separated by shale-mudstone with only thin sandstone beds. Near the base of the main quarry exposure a thickness of half a metre of shale-mudstone with thin sandstone beds shows considerable distortion (Fig. 2.1Dz). Beds have been pulled apart in places and extensive overfolds of thin beds occur with tool markings now preserved on the tops of the overturned limbs. This deposit does not appear to be a typical slump as it maintains its thickness laterally. The bed marked (x) in Fig. 2.1Dz is usual in that it retains a planar top, and appears to have been deposited whilst the distortion was taking place beneath it. Possibly this deposit was produced by gradual creep in the beds rather than by active slumping. This structure could be allied to slumping and may represent downslope movement of material, but in this case the deformation was slow and continuous

and only affected about 50cm. of sediment. The inferred palaeoslope direction obtained from this structure corresponds well with that obtained from the true slump deposits at Hawkshead Quarries nearby. On this theory the overlying sand bed was deposited after the slumping had taken place.

Structures practically identical with this deposit have been produced experimentally by Dzulynski and Radomski (1966) by the impact of a heavy suspension on a surface of horizontally bedded sediment. The deposited bed of sand from the heavy suspension has an irregular base against the underlying distorted bedding. In view of the extensive nature and consistent thickness of the structure seen in Fox Bank Quarry and the presence of a thick sand bed with a planar top above the disturbed bedding, it is considered that the emplacement of the sand bed rather than slow sediment creep or slumping was the cause of the distortion. The orientation of the folds in the experimental disturbed bedding were often parallel to the direction of flow (Ibid. p.229), as was the case in the present area.

The slump directions obtained in the Croker Hill area indicate the presence of a downward slope in an easterly direction. Current directions indicate that the main transport direction was from south to north across this slope. Care must always be taken in determining

slump directions from the appearances and trend of slump folds, and in the present area the only place where slump folds could be measured with any degree of confidence was at Hawkshead Quarries 371. Dzulynski and Walton (1965 p.195) mention that folds in slumps are often very variable in direction, and may reflect irregular local movement within the slump.



Fig. 2.1Dv

Pull-apart from slump Loc.015.

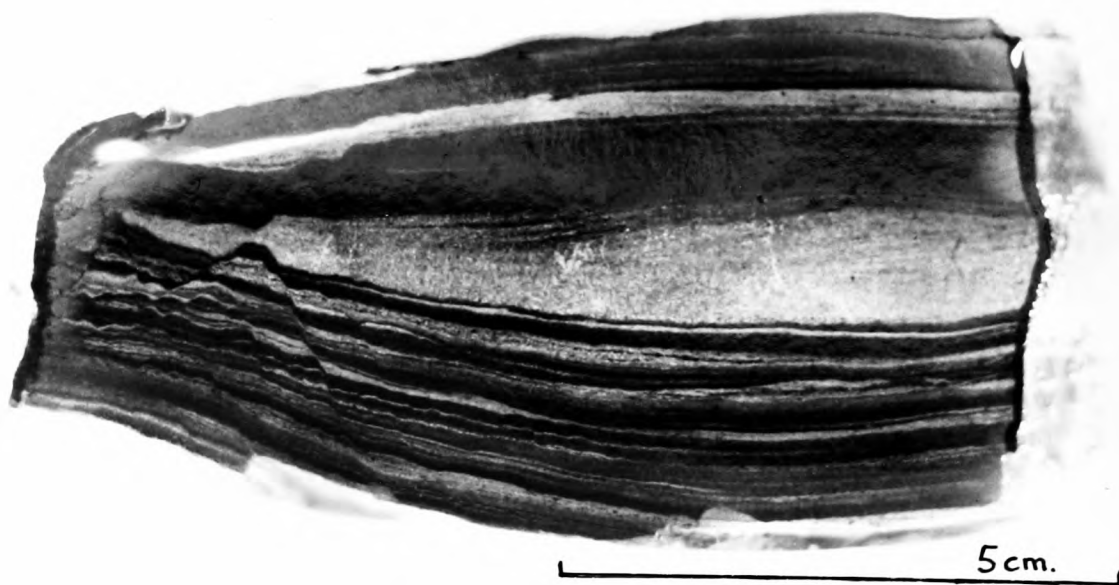


Fig. 2.1Dw

Pull-apart from slump Loc.370.

Both beds pulled apart at left of photos with microfaulting developed within the thinned parts of the beds.

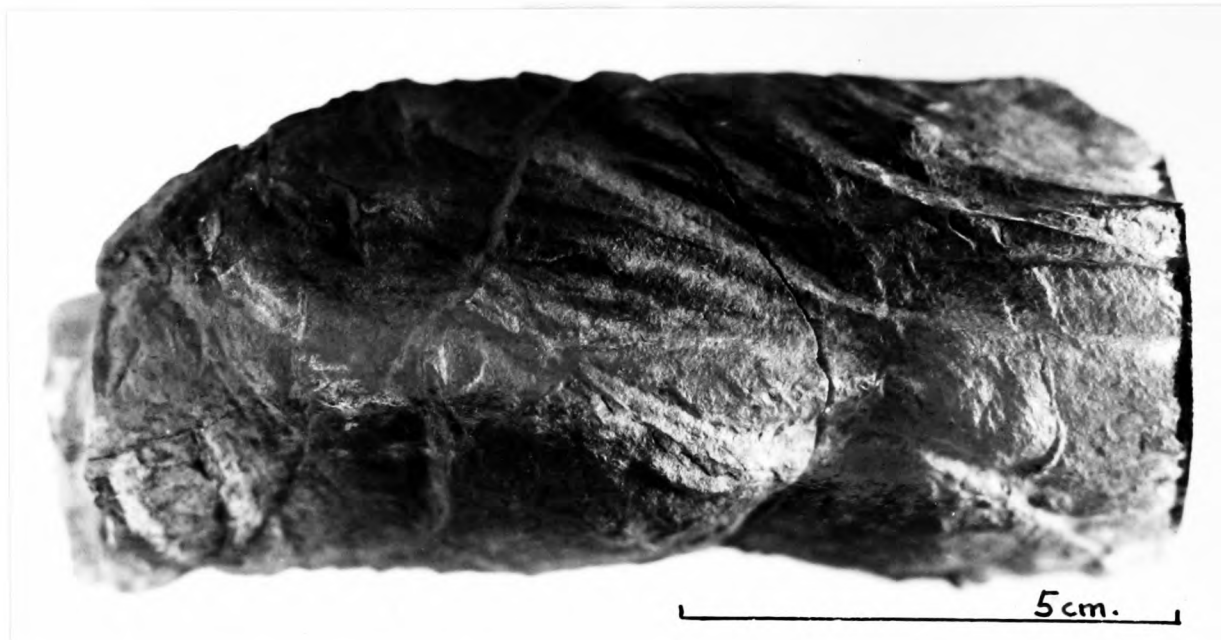


Fig. 2.1Dx

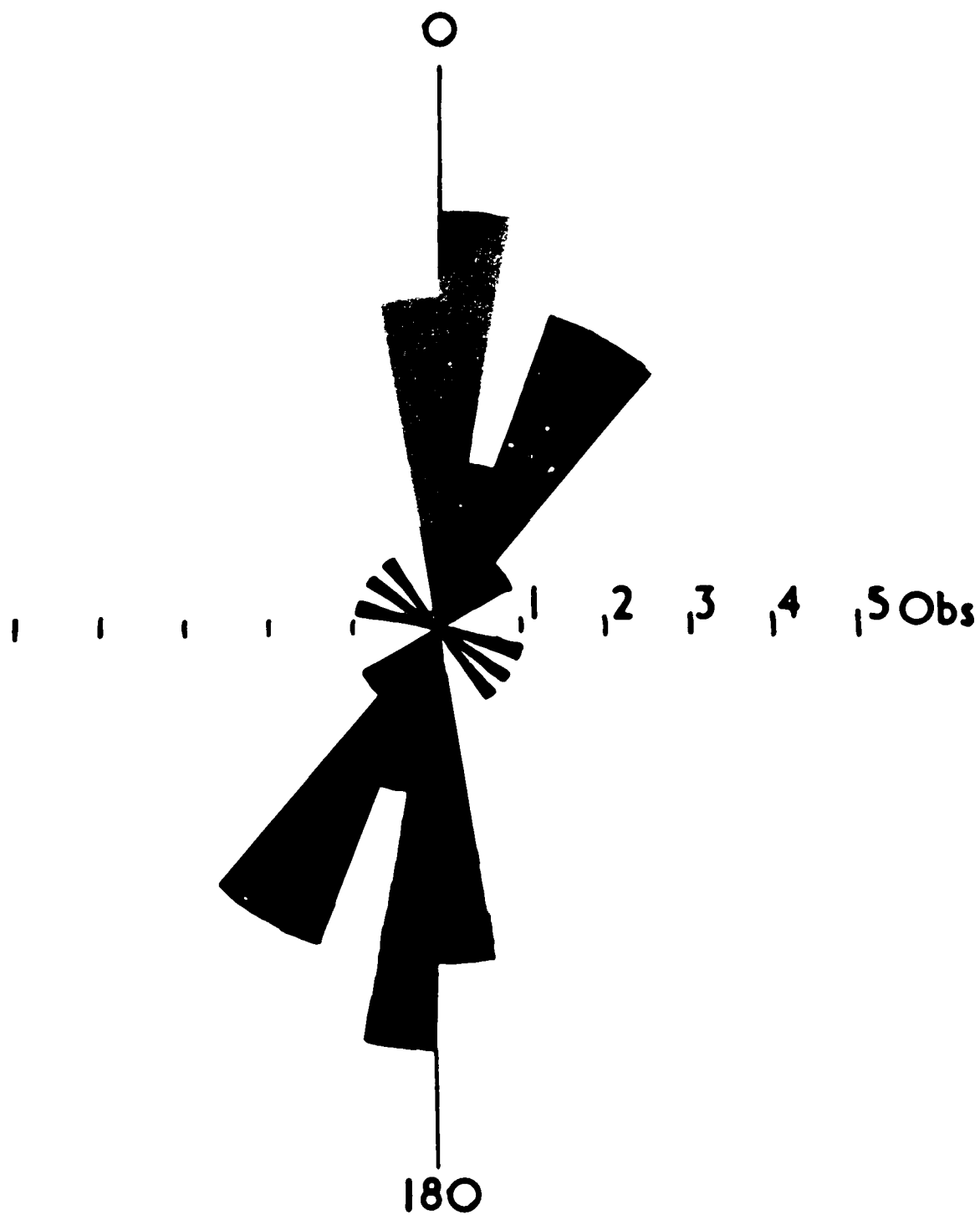
Isolated slump fold axis seen in section and plan to show pulled-apart sand laminae which form ridges on the fold axis
Loc.370.

Fig 2.1Dy

SLUMP DIRECTION

LOC 371 Hawkshead Quarries

Based on orientations of slump fold axes
and lineations

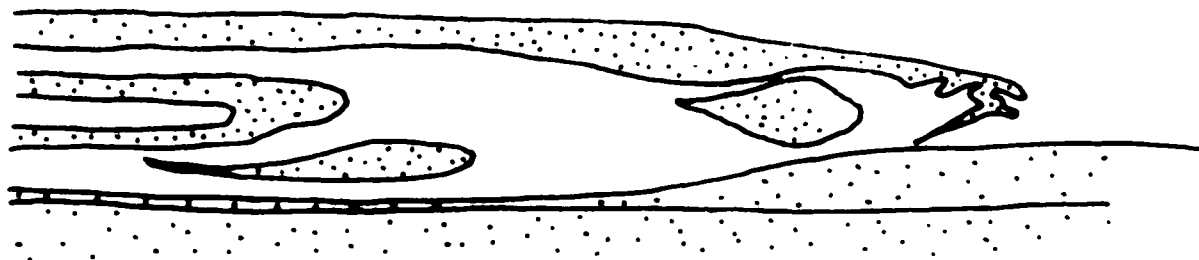
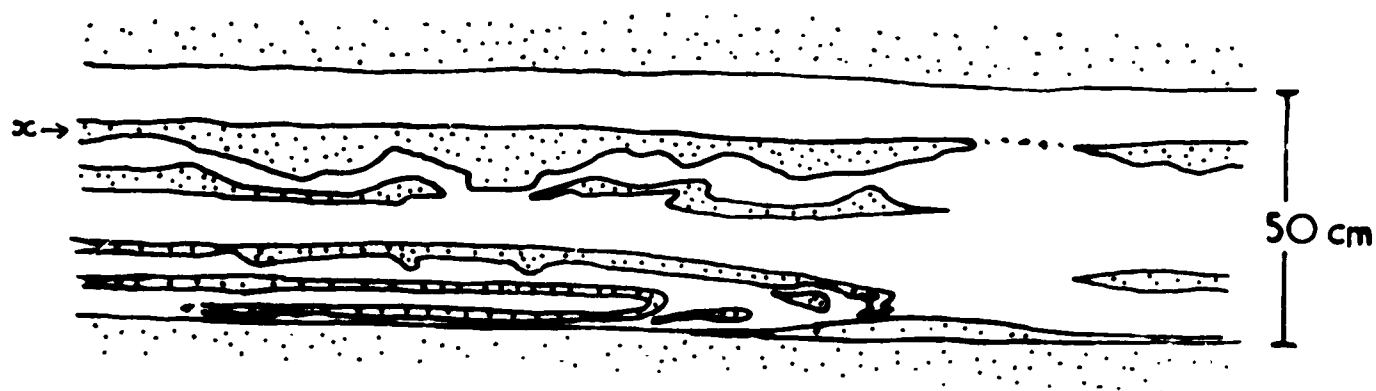


Slumping from west to east.

Fig 2-1Dz

DISTURBED BEDDING

Loc 360 Fox Bank Quarry



Detail from above.

Sand beds stippled.

Sketch of disturbed bedding in thin bedded protoquartzites
below a bed (x) with planar top and very irregular base.
The distortion was probably produced by the impact of
bed (x) on the underlying sediment.

2.1E

INTERNAL SEDIMENTARY STRUCTURES OF THE PROTOQUARTZITES.

As has already been indicated many of the proto-quartzitic beds grade upwards into a silty green shale, and in this section the structures are described which occur within a bed from its base against shale-mudstone up to its top contact with the black shale-mudstone which comprises the 'normal' basin deposit of the area. The top parts of the beds are not lithologically 'protoquartzites' since they contain a large proportion of clay material. It is however still useful to retain the term protoquartzites for this lithofacies as introduced by Holdsworth (Thesis 1963) since it is now generally used and describes the principle mineralogical features of the beds, and effectively distinguishes them from other lithofacies found in the north Staffordshire basin area.

Although, as indicated previously no rigid distinction can be made between the beds described as proximal or distal it is useful to describe separately the structures typical of each type of development.

Proximal Beds. (Fig.2.1Cb for log)

Individual beds are usually between 10 and 50 centimetres thick and consist mainly of protoquartzitic sand but occasionally contain numerous mud flakes and

sometimes rounded pebbles of quartz.

The beds are usually structureless internally and have no visible grading throughout the main part of the bed. However, if the bed contains any pebbles these are usually concentrated at the base, often in load structures. The very tops of the beds sometimes show a weak parallel lamination and here the bed may grade rather abruptly to green silty shale. The topmost laminated part is usually less than 5cm. thick.

Plant fragments, shale fragments and quartz pebbles occasionally occur apparently randomly distributed in the structureless part of the bed. Two beds at Sprink clearly showed imbrication of the constituent grains. In one case the imbricated material was small plant fragments and in the other case elongate pebbles in an unusual bed (Fig.2.1Ea) which consisted largely of pebbles of quartz arranged with a crude parallel lamination. The bed was cut parallel to the current flow direction as indicated by bottom structures on the bed and in some of the coarse laminae grain imbrication was evident in an upstream direction.

One bed was observed which was 50cm. thick of which the majority was structureless but the top 10-20cm. was distinctly coarser and showed planar cross stratification. The bed could not be traced laterally for any extent due to poor exposure but the cross-bedded top continued for

about 5 metres maintaining roughly the same thickness, and thus it does not appear to be the filling of a minor channel, but an areal feature of the bed.

These proximal type beds are usually grouped very close together in any one section and are rarely separated by more than a few centimetres. The material found between the beds consists of green sandy silt, sometimes containing isolated sand ripples and frequently very rich in plant fragments. These plant fragments may be preserved in a semi solid state due to sand filling originally hollow stems, this feature has not been seen in the distal beds.

Features of interest in the proximal beds also include the absence, or extreme rarity of burrows preserved on the bases of the beds. This is probably connected with the rarity of black shale-mudstone between the proximal beds and in which the organisms seem to have preferred to burrow. No evidence of internal burrows has been seen in any beds, and in fact no evidence of any animals or plants living in the environment of deposition of the proximal beds has been seen.

The best sections showing the proximal beds are at Shirkley, Sprink and between Endon Mill and Dalehouse Wood. The first of these localities is in Unit C, the second in Unit B with a small exposure of Unit A, and the third is partly in Unit C, but near Endon Mill at Loc. 066

one development can be shown to lie close below an E2b3 goniatite band. Thus it is slightly younger than Unit C but still in the same series of southerly derived proto-quartzites.

Thin bedded developments in proximal areas (Fig.2.1Cd for log)

The bottom structures associated with these atypical proximal developments are discussed on page 150 . Internally the beds are usually laminated and may have rippled tops against green shale. The ripples are of type 1 of Walker (1963) showing frequent truncation of the stoss sides.

The laminated beds are generally graded and some have been seen to show repeated grading due to erosion of the top of a sand bed by a later current. One bed could be split along the base of such a graded lamina and scour marks were present cutting down into the lower bed.

Sand beds may be represented by only a single line of ripples bearing tool markings on their bases, Fig.2.1Eb.

Internal Structures. Distal Beds. (Fig.2.1Cc for log)

The distal protoquartzite beds exhibit a variety of internal structures which always follow the sequence described by Bouma (1962 p.49). Thus the upward sequence of structures in a bed follows the scheme:- a, massive division; b, laminated division; c, ripple division;

d, fine laminated division, and e, graded mudstone division. Bouma (1962) originally called these divisions 'intervals' the term division being introduced by Walker (1965) and used by Walton (1967). It is unusual for all these divisions to be seen superimposed in any one bed, but the upward sequence of structures is usually in agreement with this ideal pattern. Exceptions were often observed in the case of the ripple interval when the sand bed had a sharp top against shale-mudstone and it is thought that at least two main types of rippling are present and that they have different origins. Thus the beds with gradational tops are considered first and then the beds with sharp rippled tops.

Distal Beds with Gradational Tops.

a, Massive Division.

The massive division, when present, is invariably at the base of the bed and may consist of from a few centimetres to about 30cm. of sand in which there is no visible lamination. Grading is sometimes apparent in this division with all the coarser grains concentrated at or very near the base of the bed. The sand material making up the beds is often rather uniform in grain size and as it consists mostly of quartz which has undergone secondary silica cementation grading is frequently absent, or at least not apparent. Occasional plant fragments

and shale pellets are seen in the massive interval and they may lie at high angles to the bedding but are usually roughly parallel to the bedding.

b, Lower Division of Parallel Lamination.

There is a gradation from beds with a massive division at the base to those where the base is laminated with the basal lamina 1-3cm. thick. Lamination in the protoquartzites is generally picked out by partings rich in plant fragments or clay material. Generally the laminae are developed in protoquartzitic sand. The thickness of the laminae varies from 2cm. down to a few millimetres, but laminae of 5-10mm. thickness are most commonly met with. The thickness of the division ranges from a single lamina up to about 10cm..

Payne (1942) using the term lamination, suggested 1cm. to be the upper limit of the term, but in discussing these beds this is inconvenient since many laminae in the b division exceed 1cm. in thickness. The concept of the laminated division is well known and it seems unnecessary to apply limits to the thicknesses of the laminae.

c, Ripple Division.

As already mentioned two types of ripple lamination are present in the protoquartzites. The type found within the protoquartzite beds corresponds with Type 3

of Walker (1963) and appears to have been formed during conditions of net deposition giving ripples with continuous laminae which are not truncated (Fig.2.1Ec). The wavelength of the ripples is generally 5-10cm. and amplitude up to 2cm. The ripples come into being gradually, developing from the underlying parallel lamination. Upwards the ripples may also die out gradually into finer laminated sediment. These ripples have not been observed in plan view over any large surfaces and their exact shape cannot be determined. They are certainly not straight crested and probably most closely resemble linguoid ripples. This ripple division is frequently not developed in the protoquartzite beds, or at least cannot be detected in the field. The ripple division has not been observed to exceed 5cm. in thickness.

d, Upper Division of Parallel Lamination.

In the d and e divisions the beds cease to be protoquartzitic in composition. Green silty laminae first appear alternating with sandy laminae a few millimetres thick. Cementation of the quartz grains is poor or absent at this level since there is a much larger proportion of clay present than in the a, b and c divisions. The grain size of the quartz material is visibly reduced relative to the protoquartzitic part of the bed and this reduction in grain size frequently appears to be abrupt at the top of the b or c division. The poorly cemented

nature of the d division makes collection of specimens difficult and in weathered outcrops this division and the e division are difficult to observe, although stream washed sections only a few yards away may clearly show their presence. When it can be observed the d division is usually thin, consisting of a few quartz rich laminae separated by silty green shale.

When division c is absent there is no way of certainly distinguishing the b and d divisions. The base of the bed usually consists of thicker laminae than the top and there may be a marked break where the bed becomes richer in clay upwards. In the field it is not possible to divide such beds into b and d divisions. It is possible to note the general thickness of the laminae and it has been found that beds with laminae generally greater than 5mm. thick are usually protoquartzitic and resemble the b division. Beds which contain a high proportion of clay material are generally not cemented and have laminae of sand usually less than 5mm. thick and are considered to represent the d division.

e, Graded Pelitic Division.

The recognition of this division is not easy when exposures are weathered. It is best seen in stream washed sections where it shows up as a green band at the top of a bed with a sharp contact against the overlying black shale-mudstone which forms the normal basin sediment.

This pelitic division is often visibly graded and contains quartz grains of fine sand grade. The bulk of the interval consists of silt or clay and flakes of mica are often abundant. In the distal areas many beds occur consisting only of e or d-e units and since they are only poorly lithified and are not easily studied in the field. The thickness of this division varies from a mere parting up to 8 or 10 centimetres but the usual range of thickness is from 1 to 5cm.

General Considerations.

The five different divisions in a bed described above form an ideal sequence which was rarely seen in the field. By far the most common are 'base cut-out' beds (Bouma 1962, p.50) in which one or more of the lower divisions are absent. This gives sequences of divisions within a bed as follows bcde; cde; de; e. All these combinations have been observed in the field in the present area.

Another type of bed that has already been mentioned is one where the c division is absent and the sequence abde or bde is present. The difficulty in separating the b and d divisions in the absence of the c division has already been mentioned, and it might be of more practical use to describe them as being 'laminated sand with graded pelitic top'. They could be referred to

notationally as ab-de and b-de to indicate that divisions b and d cannot be separated.

Beds with Sharp Rippled Tops.

Throughout the area, within developments of the protoquartzites, beds occur that are protoquartzitic and possess a sharp upper surface against black shale-mudstone of basinal type. The d and e divisions are thus absent. The bases of the beds are sharp and do not differ from the other beds to be found in that they have the same assemblage of bottom structures.

Internally there is an important difference between these beds and the beds containing a c division and having gradational tops. Whereas truncation of ripple laminae was absent or rare in the c division and the ripple lamination conformed to Type 3 of Walker (1963). In these beds the ripples frequently have erosive bases and truncation of ripple laminae is of frequent occurrence (Fig. 2.1Ed). These ripples correspond to Type 1 of Walker. It is clear that these ripples have not arisen during times of net deposition, but are the result reworking of previously deposited material. The current that caused the rippling has carried away any fine material that was present and remoulded the top of the sandy parts of the beds into a ripple form. The time at which this reworking took place is open to question

and is discussed later when the origins of the proto-quartzites are considered. Study of the current directions indicates by both bottom structures, and by ripples, currents coming from the south.

Graded Bedding.

The grain size distribution in the protoquartzites has not been studied in detail for the reasons outlined in the petrography section, (i.e. recrystallisation of quartz, replacement by siderite, etc.) but overall grading is often visible in the field and it is most often seen at the tops of the beds, where dark clay and silt material becomes abundant and detrital mica is concentrated. The lower, protoquartzitic parts of the beds often do not appear to be graded probably due to the absence of notably coarse grains in the original material of the turbidity current. Some beds contain coarse grains concentrated at the base and show normal grading, and a few beds have been seen which show multiple grading (Ksiazkiewicz 1954) probably due to the amalgamation of deposits due to two turbidity currents. Similar features are described by Walker (Thesis 1964) in the north Derbyshire Namurian. Frequently at the base of a bed containing a range of grain sizes suitable for showing grading the basal few millimetres show reverse grading before normal grading sets in. This has been termed pen-symmetrical grading

(Ksiazkiewicz 1954, p.406). Discontinuous grading with an abrupt change in grain size within the bed is apparently common in the area studied, there often being a marked break below the d-e divisions which are poorly cemented and the rest of the bed which is generally silica or siderite cemented. It is probable that this break marks the incoming of a considerable amount of clay matrix to the quartz which has inhibited cementation rather than any marked change in the size of the quartz grains.

Convolute lamination.

Convolute lamination has been observed associated mainly with the finer grained portions of the beds where a high proportion of silt appears. Few examples have been seen, probably because beds of suitable composition and grain size are not common, and because these beds are not always well cemented. Fig.2.1Ee illustrates convolute lamination typical of that seen in the area.

Structure due to postdepositional sediment movement. Fig.2.1Ef

In this bed the base is laminated and the undersurface bears groove and prod moulds. There is a sharp division between the lower laminated part of the bed and the upper disturbed part. The junction is marked by a concentration of plant debris which has acted as a plane along which movement has taken place. Above this plane all original

structure is destroyed. Parts of the overlying shale-mudstone have been incorporated in the top of the bed giving it an irregular top and indicating that the movement was postdepositional.

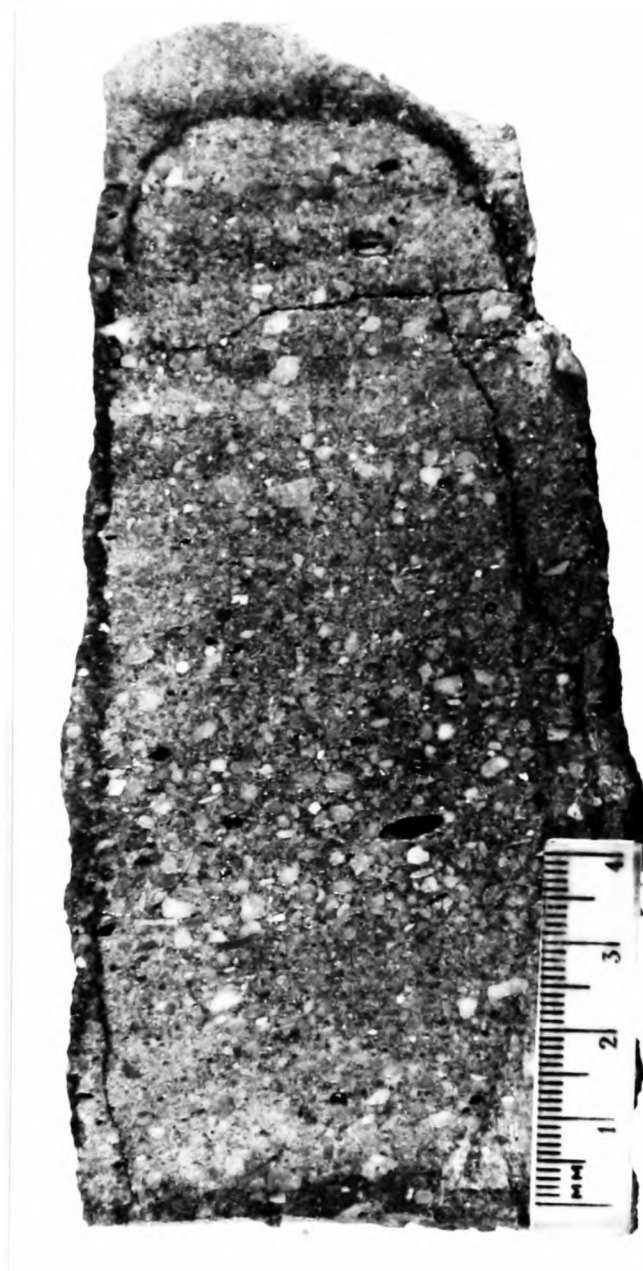


Fig. 2.1Ea

Very coarse proximal protoquartzite Unit B. Loc.073.



Fig. 2.1Eb

Section of thin bed from proximal area Loc.054 consisting of a single ripple which is cut through at left of photo by a scour with coarser fill from an overlying bed.

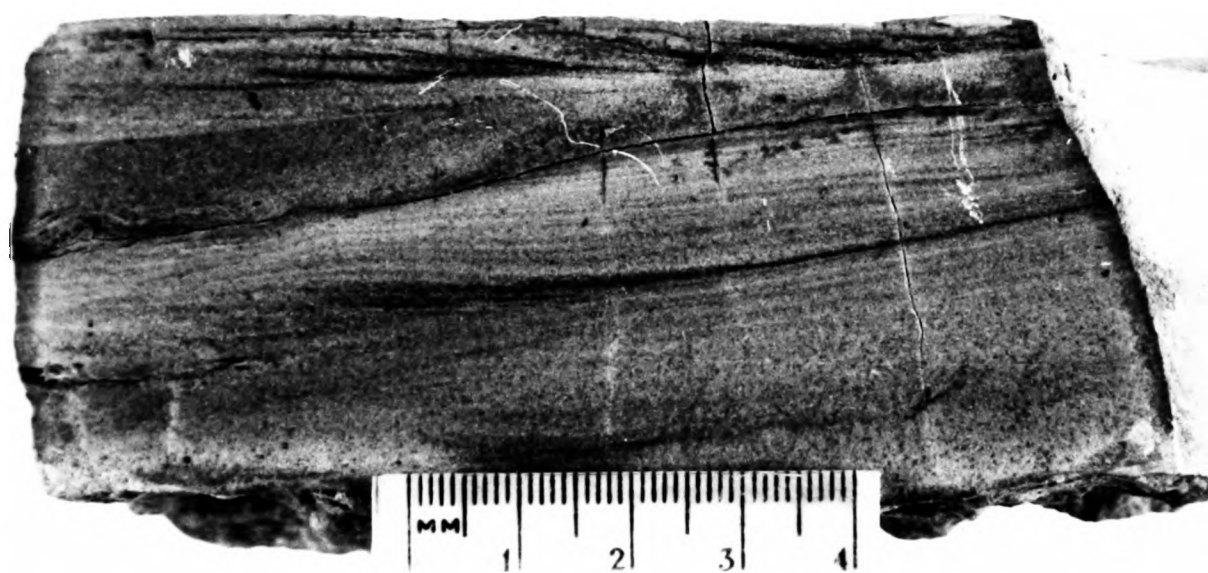


Fig. 2.1Ec

Ripple lamination typical of division c of the protoquartzites with laminae generally continuous, and truncation of laminae due to erosion absent.

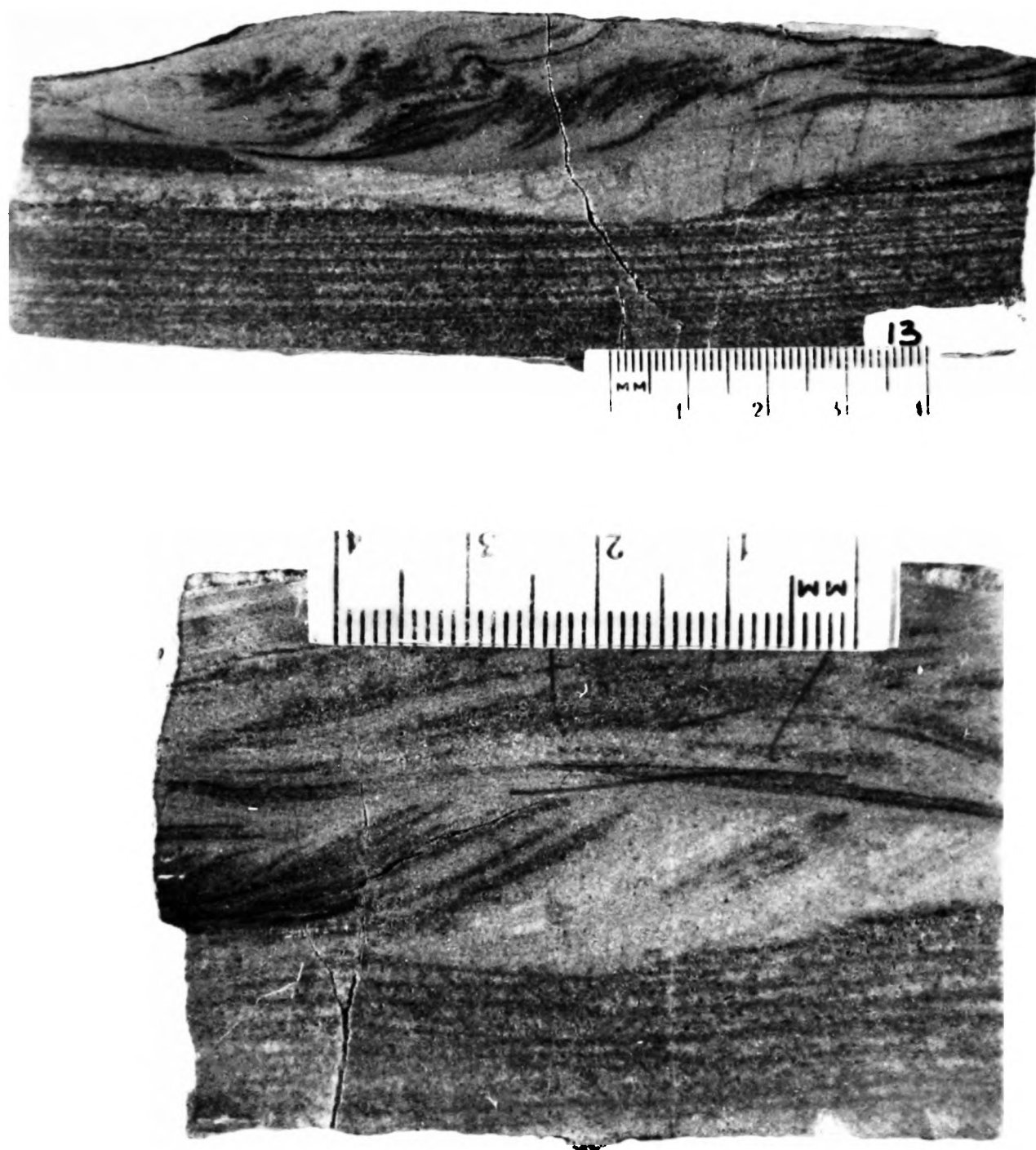


Fig. 2.1Ed

Protoquartzite bed with laminated base and rippled top. The ripples have an erosive base and result from reworking of previously deposited material. Truncation of ripple laminae frequent. Base of Unit C. Loc.351.

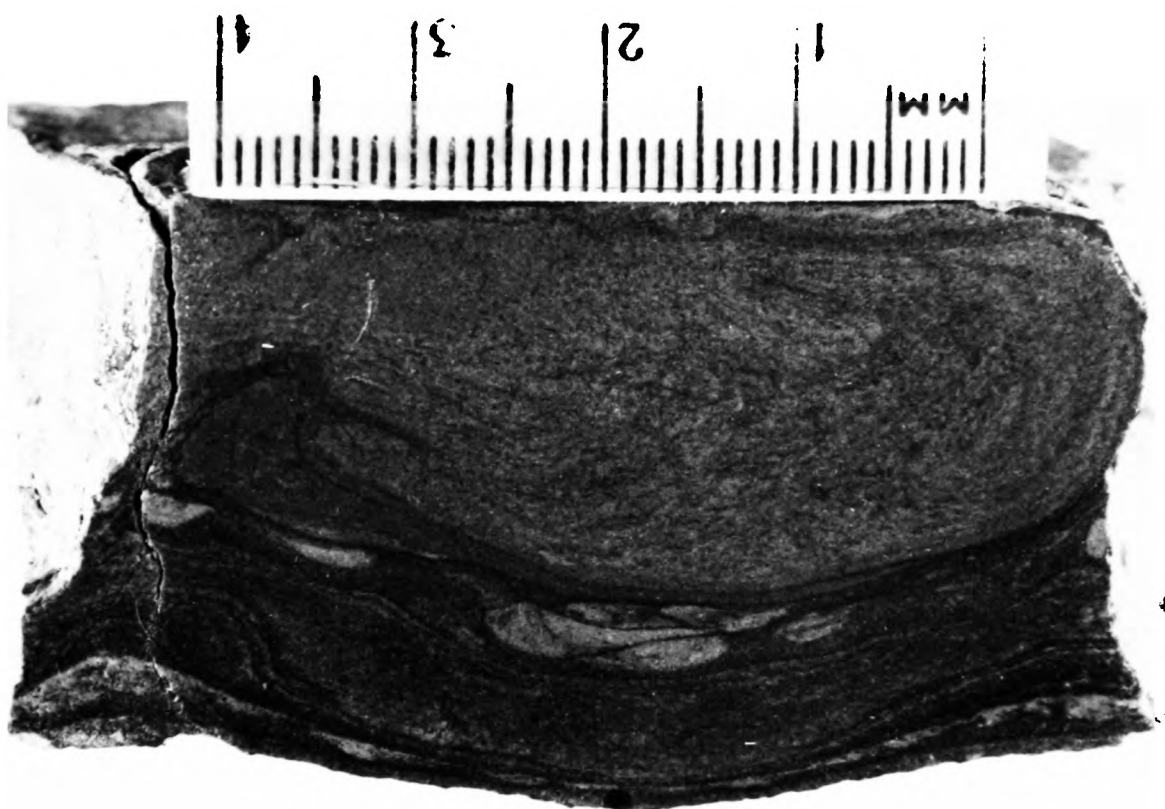


Fig. 2.1Ee

Convolute lamination, both examples from Unit A. Loc.102.

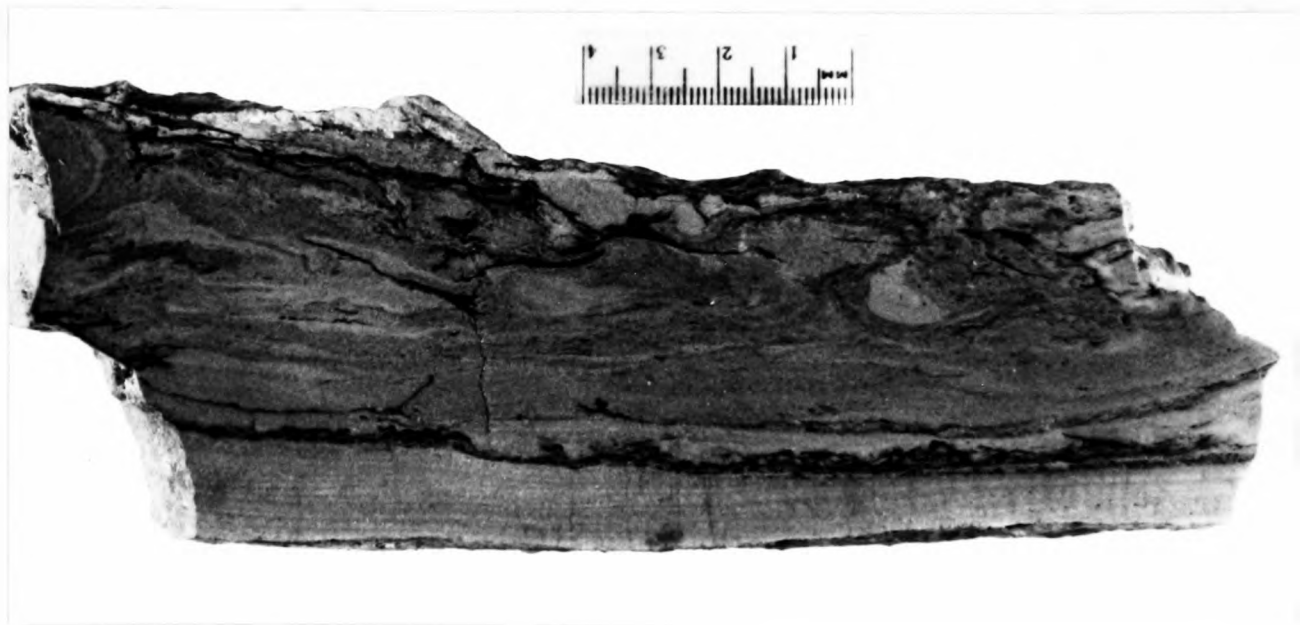


Fig. 2.1Ef

Structure due to post-depositional movement. Loc.110 Unit B.
Bed with laminated base and distorted top incorporating part
of the overlying pelagic shale-mudstone.

2.1F

PETROGRAPHY OF THE PROTOQUARTZITIC TURBIDITES.

GENERAL.

The petrography of the turbidites has not been studied in detail for several reasons. The object of such a study would be to equate functions such as grain size, sorting and proportion of clay matrix to the individual depositional phases of the turbidite. The factors which make such a study impossible on the present material are as follows:-

- 1) Partial replacement of many beds by siderite, matrix apparently being replaced more easily than detrital sand grains.
- 2) Extensive secondary silica cement which may constitute nearly 20% of the rock, and make original grain sizes and shapes impossible to determine. This also alters the original ratio of matrix to sand grains.
- 3) The presence of all stages of disintegration of shale pellets. It is therefore difficult to decide if a particular fine grained patch represents original disseminated fine grained material or a single shale pellet squeezed between its surrounding sand grains during compaction.

The combination of these three factors gives a resulting rock which is probably far removed from the

originally deposited sediment, and thus it would be dangerous to draw any conclusions on the character of the original sediment. It has already been mentioned that it is only the basal parts of the turbidites which are protoquartzites (Pettijohn 1957 p.366). The protoquartzitic part of the bed corresponds roughly to the a, b and c divisions of the turbidite, and is generally hard and silica cemented. The upper parts of the beds are finer grained and much richer in clay matrix and are generally less well cemented. The lack of cementation in the d and e divisions makes collection and sectioning of specimens difficult. The tops of the beds are also rich in carbonaceous debris which is uncommon in the lower parts of the beds except on isolated parting planes.

The petrology of the Minn Beds of Evans et al. (1968) has been described by Harrison (in Evans et al. 1968 p.83). Most of the rocks studied were probably of protoquartzite lithofacies but since the authors do not recognise the calcareous siltstones as a separate lithofacies then some of these beds may also be included. In general the petrography is in agreement with that described here, but there are two exceptions. The first is a record of glauconite which is unusual for the basin environment. The second is the absence of siderite from the list of secondary minerals, when it has been found frequently by Holdsworth (thesis 1963) and the present writer. Harrison

records calcite and ankerite and this is probably due to the inclusion of some calcareous siltstone beds in his samples.

The protoquartzites described correspond with Lithofacies 2 of Holdsworth (thesis 1963 p.231) and the features described briefly below are in agreement with those noted by Holdsworth for these beds. Photomicrographs of typical protoquartzites are shown in Figs.2.1Fa-c, and illustrate increasing proportion of matrix with decrease in grain size.

MAJOR CONSTITUENTS.

Quartz.

Quartz is the most abundant detrital mineral present. Grains may exhibit ordinary or strained extinction and in the coarser beds the proportion of polycrystalline detrital grains may be in excess of ordinary grains. Contacts between individual crystals in polycrystalline grains may be sutured, suggesting a metamorphic origin. Regularly oriented inclusions in quartz grains are rare, the majority of grains being reasonably clear. Some grains of finely crystalline quartz are possibly derived from rhyolites or other acid volcanic rocks. Occasional grains of detrital chert are also found in the protoquartzites.

The secondary silica which is so important in the protoquartzites occurs in two forms. Usually it is present

as optically continuous growths on original detrital quartz grains and frequently this gives straight boundaries between grains (Fig.2.1Ff). In some cases the original grain outlines can be seen but often the original shape of the detrital grain cannot be determined. The second form in which secondary silica can occur is as chert filling voids between grains or possibly replacing matrix. Chert is often seen crystallized within the matrix material.

Matrix.

Next in importance to the detrital quartz is the matrix material. This comprises several minerals which often cannot be distinguished using the microscope. Clay minerals are probably the most important constituent of the matrix and these have been identified by X-ray diffraction and consist of kaolinite, chlorite, muscovite (illitic) and mixed layer mica/montmorillonite (see section 1.4). Using the microscope only the kaolinite and micaceous material can be identified with certainty. The micaceous material is certainly largely detrital but some of it may be diagenetic in origin (Holdsworth thesis 1963). The kaolinite is present both as very fine grained material and also as vermicular growths which often occur in patches. These patches may represent alteration of original detrital grains such as feldspar or they may have grown in the sediment during diagenesis. The partial alteration of the K-bentonites (section 1.4) to kaolinite

indicates that kaolinite was probably stable in the protoquartzite depositional environment.

Clay minerals make up a large proportion of the matrix, but also present is fine grained quartz and some carbonaceous material. The carbonaceous material is more abundant at the tops of the beds where it occurs as large fragments, but it is also present in a finely divided state within the matrix.

Secondary quartz or chert is frequently seen within the matrix material, often extending from the surrounding quartz grains. It is not possible to determine whether the secondary quartz actually replaces matrix material or merely fills original voids. As can be seen from the modal analyses, secondary quartz is abundant in the protoquartzites.

Shale Fragments.

Elongate shale pellets seen in thin section are similar to the shale-mudstones occurring between the turbidite beds, and were probably eroded from the substrate by the passage of the turbidity current. A gradation is seen in thin sections between obviously discrete shale fragments and general matrix material. After deposition the finest of the shale fragments were squeezed around the sand grains and are now indistinguishable from original fine grained matrix.

Kaolinite.

Kaolinite occurs in three ways within the proto-quartzites. It may be finely disseminated within the matrix, but detectable by staining, and possibly detrital in origin. It also occurs as patches of well crystallized vermicules which are clearly not detrital and may represent in-situe replacements of feldspar (Holdsworth thesis 1963 p.247), although no feldspar is now present in the rock. The third mode of occurrence is along the cleavages and at the ends of detrital mica flakes. In this case the kaolinite appears to be derived from the mica. Holdsworth (thesis 1963) noted vermicular kaolinite crystals with some layers apparently of mica which he considered was forming from the kaolinite. Both processes may have occurred.

Mica.

Muscovite mica occurs as typical detrital flakes some of which appear to be in the process of alteration to kaolinite. No biotite has been seen by the author but Holdsworth (thesis 1963) records rare altered grains. It is possible that some of the white mica is diagenetic in origin, and this will be considered later.

Carbonaceous Material.

Plant fragments are abundant in the turbidites and tend to be concentrated on parting planes. They are rare

in the a, b and c divisions apart from on parting planes, but are frequent within the d and e divisions where carbonised plant material may form over 10% of the rock (Fig.2.1Fe).

MINOR CONSTITUENTS.

Feldspar.

No K-feldspar has been seen in any of the proto-quartzites although patches of vermicular kaolinite may represent altered K-feldspar grains. Any K-feldspar present in the original sediment was probably mostly removed by weathering before reaching the depositional area. Plagioclase is present in trace amounts - generally one or two grains in a slide and is remarkably fresh and suitable sections show it to be oligoclase or albite. Holdsworth (thesis 1963 p.234) also found plagioclase of this composition in these beds.

Zircon.

Zircon is present in all slides, usually as typical rounded detrital grains.

Tourmaline.

Tourmaline is also present in all slides and is less well rounded than the zircon, it is usually pleochroic in green and yellow-green.

Garnet.

No grains of garnet have been seen by the author, but Holdsworth (thesis 1963 p.234) records rare highly altered grains.

Ore Minerals.

Pyrite is occasionally present as cubes of secondary origin in the protoquartzites, and is often present as fine grained framboidal aggregates (see Holdsworth thesis 1963 p.246). The framboidal pyrite is associated with the matrix and thus is more frequent in the clay rich tops of the turbidite beds.

MODAL ANALYSIS.

General techniques.

Modal analyses of the protoquartzites are difficult to express with accuracy for a variety of reasons. The most difficulty was experienced in distinguishing detrital and secondary quartz, and the figures given for secondary quartz should be treated with caution since in some slides the secondary quartz is more easily distinguished. In the presence of extensive secondary quartz it is difficult to distinguish polycrystalline grains. The percentages of polycrystalline and secondary quartz should be regarded as minimum values.

The term 'matrix' includes all fine grained material,

mainly clay and fine carbonaceous material, acting as matrix and in which individual minerals cannot be determined while point counting. There is a gradation between 'matrix' and 'shale fragments' and it is probable that some material counted as matrix was deposited as a shale fragment which was subsequently distorted. Thus the values quoted for 'matrix' may be higher than the actual original amount of fine grained matrix. Conversely, the extensive secondary silica cement encroaches areas that were either originally void or matrix, and thus the matrix % may be low in relation to that present when the bed was deposited.

Cummins (1962) suggested that the matrix of greywackes was largely secondary in origin being derived from alteration of labile grains. The absence of unstable grains in the protoquartzites could indicate that they have been altered to matrix, but this could equally well have happened before deposition. The extensive secondary quartz present indicates that matrix has probably been decreased by diagenesis rather than the converse.

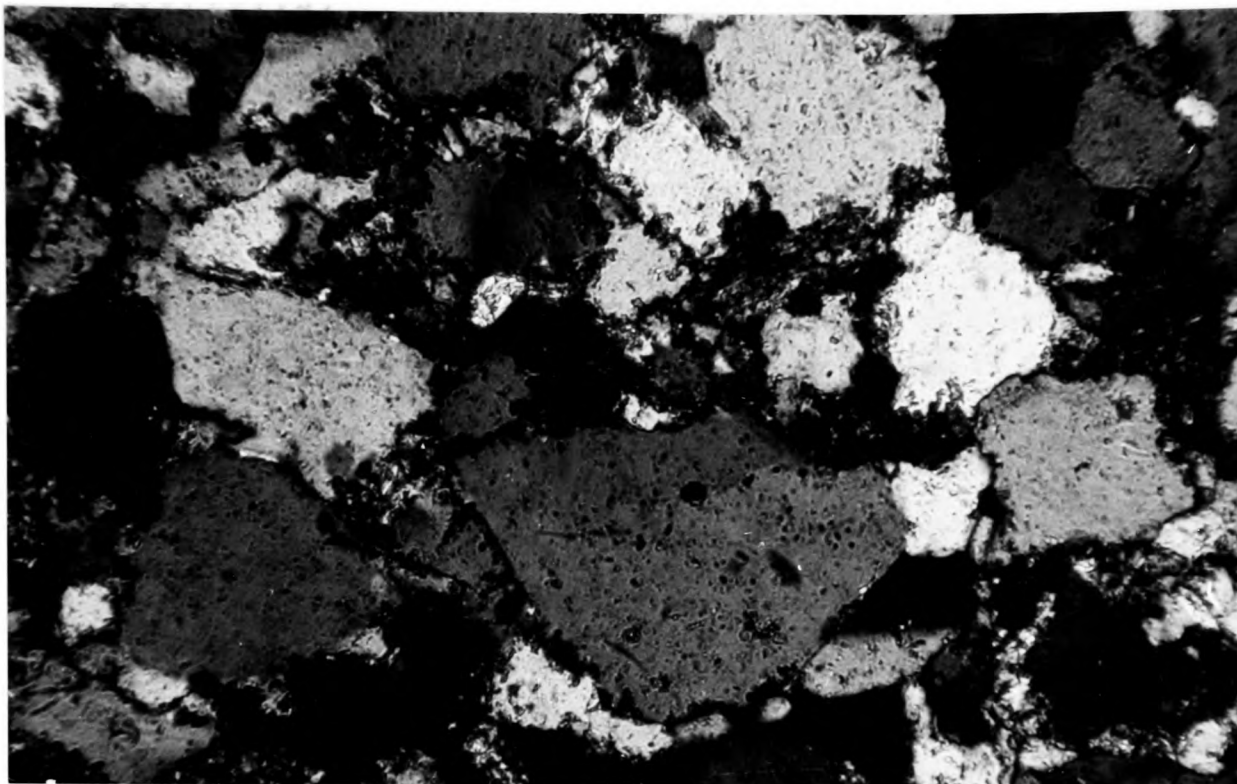
Most of the slides examined contain a significant proportion of holes (up to 4.6%). These voids may be natural but some certainly represent lost material during slide making, and in some sections it is clearly the soft shale fragments and patches of vermicular kaolinite that are lost. Thus these two constituents may be low in the modal analyses.

Modal analyses were all based on 500 points. Increased accuracy could possibly be obtained by counting more points but this would be offset by the personal observer factor which in the case of these rocks would probably be large due to the difficulties already mentioned. Modes are presented below for representative samples of the lower protoquartzitic parts of the beds (a,b,c divisions) and for the finer grained and less well cemented upper portions of the beds (d,e divisions).

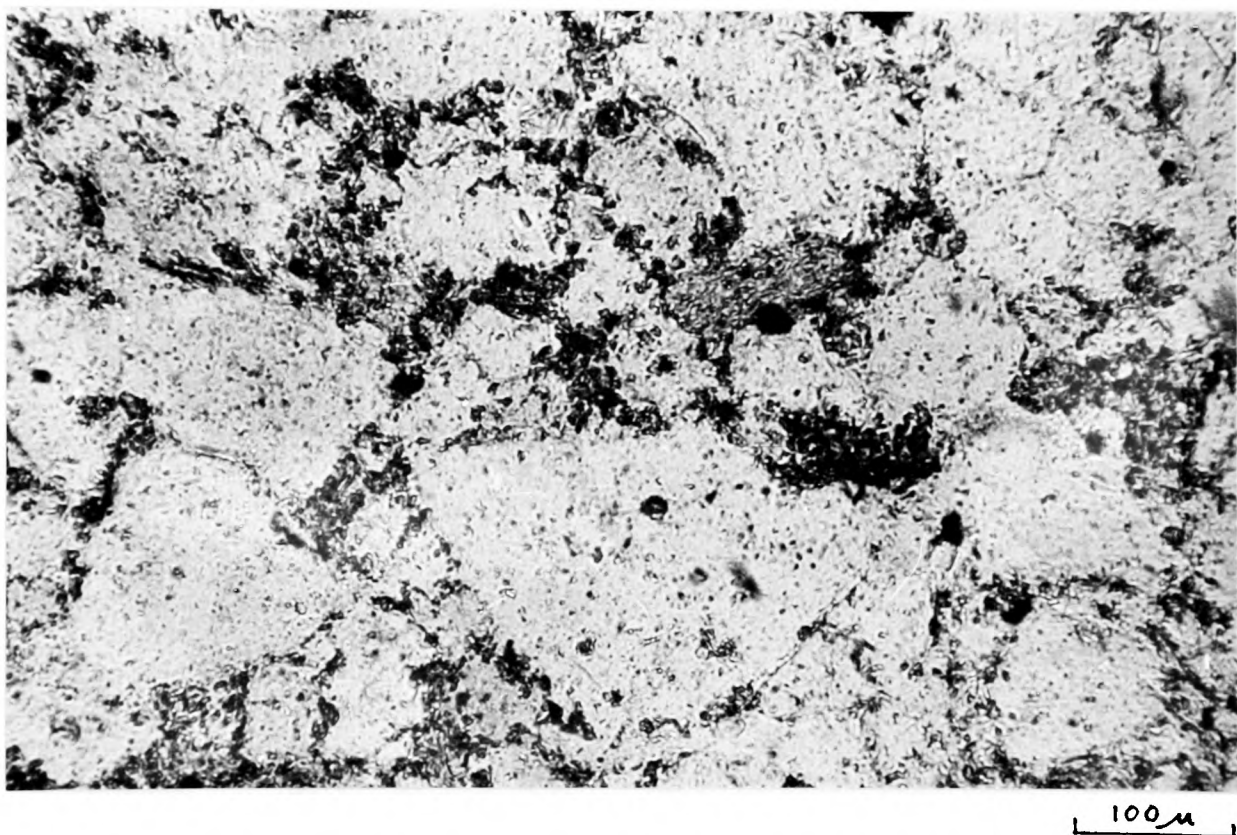
The modal analysis presented in Fig.2.1Fd show clearly the quartz rich nature of the detrital fraction and the importance of the secondary chemical cement. These features correspond to those of protoquartzites (Pettijohn 1957 p.316-7). The virtual absence of feldspar in these rocks was stressed by Holdsworth (thesis 1963 p.237) who considered 'Non-feldspathic protoquartzite' a suitable term for these rocks.

The upper portions of the beds of which modal analyses are presented in Fig.2.1Fe show an increase in fine grained material. Quartz is represented by finer grain sizes and the proportion of matrix is greatly increased. As Holdsworth (thesis 1963 p.240) points out, this leads to a greater proportion of matrix material being present and as matrix increases a whole series of petrographical names could be applied to the rocks. Some distinctly unsorted rocks with more than 15% matrix deserve the term

Fig. 2.1Fa



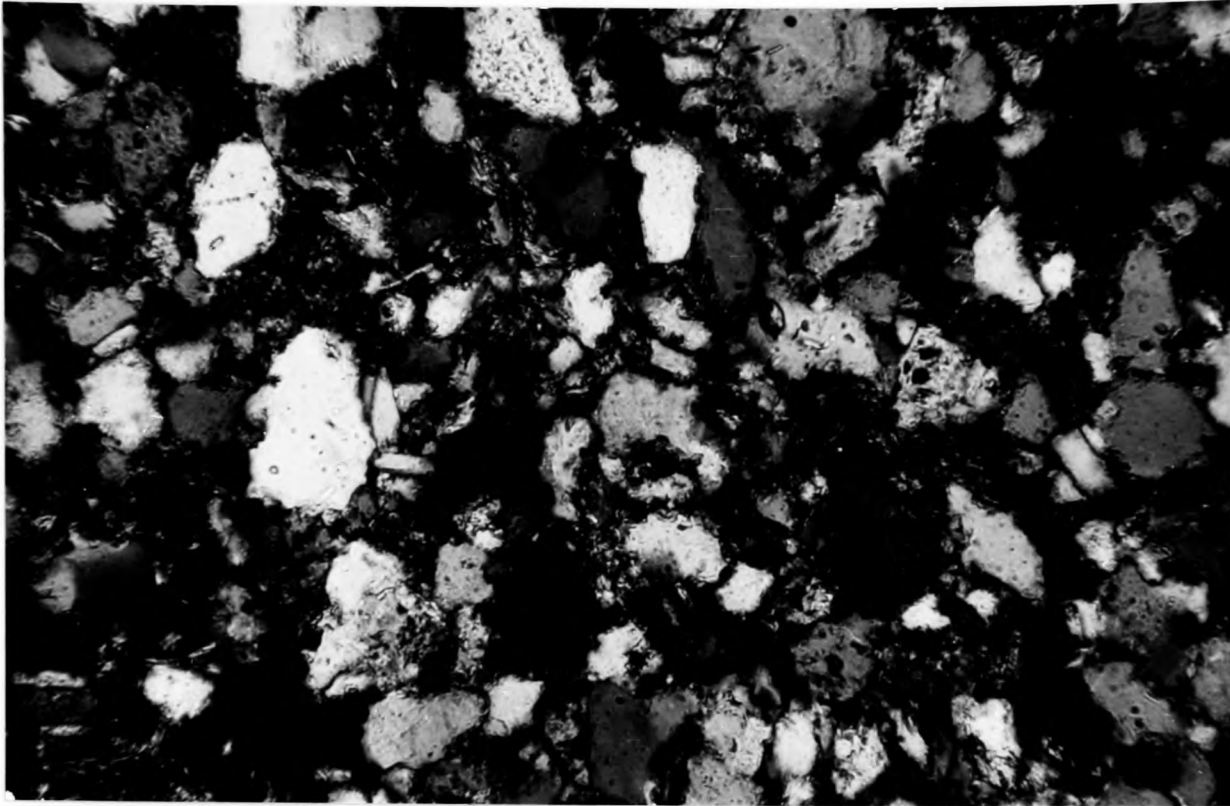
Crossed polars.



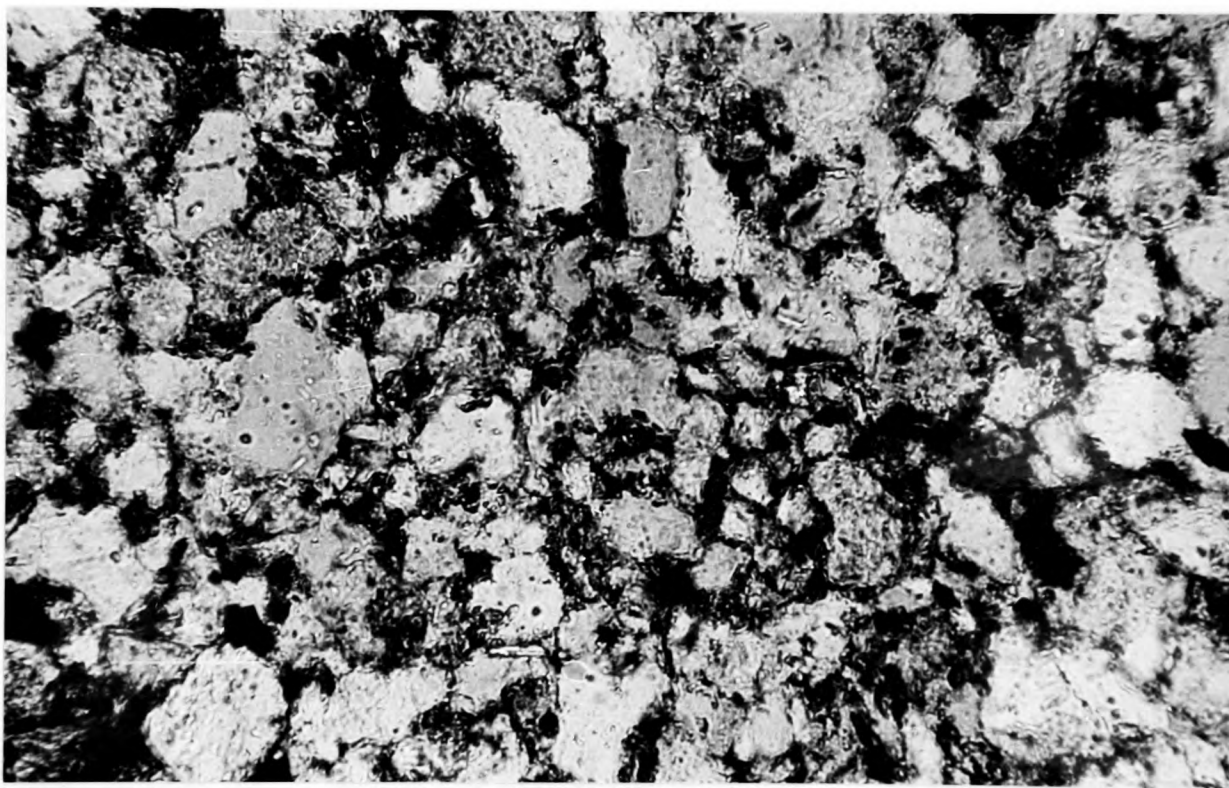
Typical protoquartzite showing the generally low proportion of matrix material present. Slide 107.1. Unit C. Loc.107A.

Figs. 2.1Fa-c are on the same scale to illustrate the increase in proportion of matrix material with decrease in grain size.

Fig. 2.1Fb

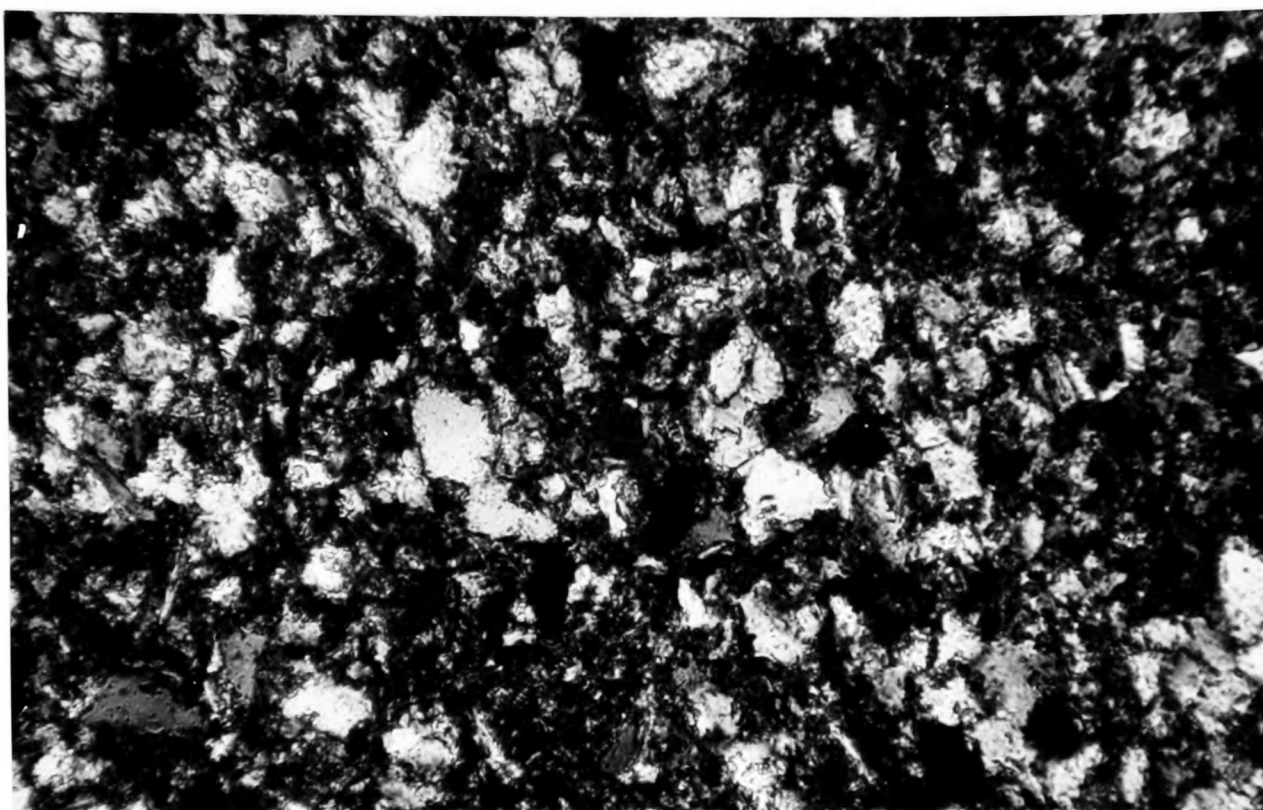


Crossed polars.

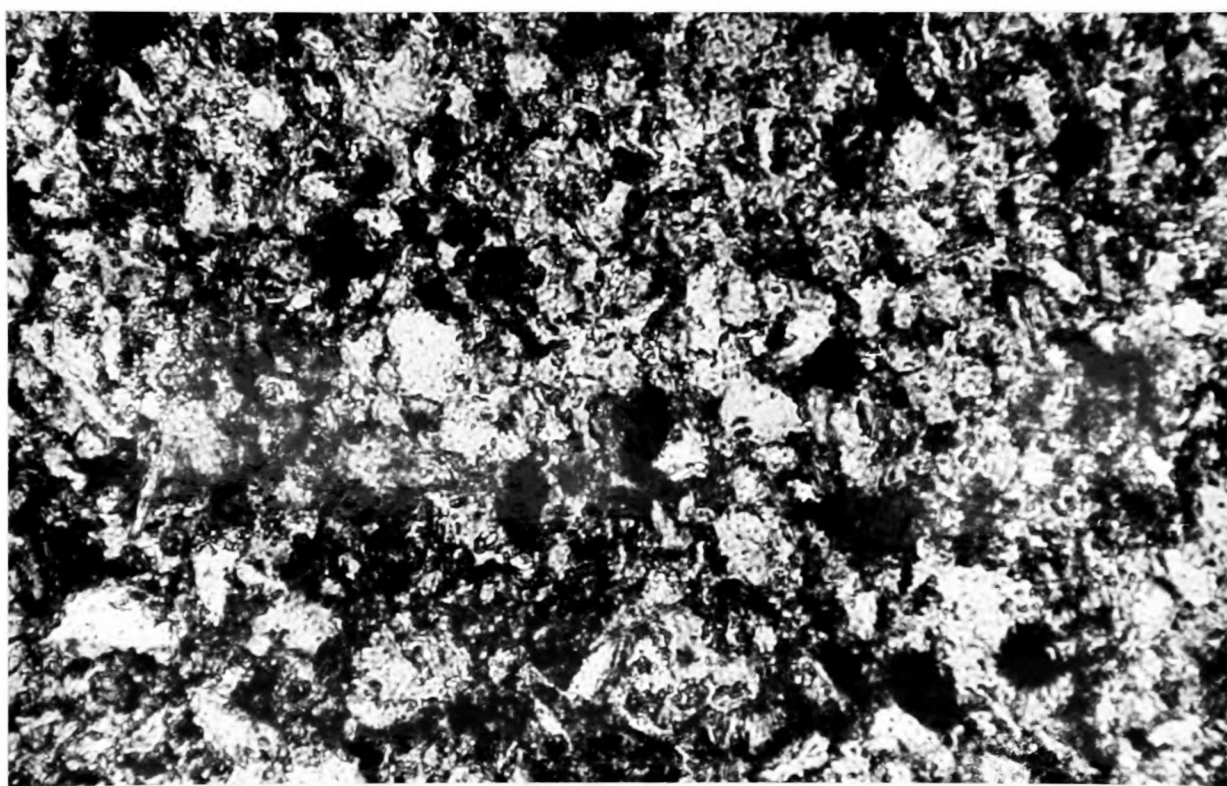


100 μ

Protoquartzite of medium grain size showing increase
in proportion of matrix material. Same scale as
Figs. 2.1Fa,c. Slide 009.3. Unit C. Loc.009.



Crossed polars.



100 μ

Thin section of the d division of a protoquartzite bed to show the approach to greywacke texture with increase in matrix material at the top of the bed. Same scale as Figs. 2.1Fa-b. Slide 102.1. Unit A. Loc. 102.

Fig.2.1Fd

MODAL ANALYSES OF PROTOQUARTZITES.

Component	Spec. No.	009.2b	072.2	401.1	401.2	073.2A	072.1A
Quartz		60.2	54.01	69.6	70.2	36.6	30.9
Multigranular Qtz.		8.4	13.87	N/D	N/D	34.9	43.3
Secondary Quartz		14.4	11.86	17.2	18.6	4.0	4.3
TOTAL QUARTZ		83.0	79.74	86.2	88.8	75.5	78.5
'Matrix'		11.2	8.76	7.8	3.3	7.5	9.4
Shale fragments		1.4	2.74	1.4	1.2	6.6	2.8
Kaolinite							
verm. patches		-	4.38	-	-	1.2	2.6
Mica		1.4	-	0.2	0.4	-	-
Others		1.6	.55	-	-	-	-
Volcanic fragments		-	-	-	3.3	6.0	4.3
Voids		1.4	3.83	4.6	3.0	3.3	2.4
		c	Proximal	b	c	Proximal	Turbidites
		division	Turbidite	division	division		
		ripples		laminae	ripples		

Fig.2.1Fe

MODAL ANALYSES. Fine grained tops of protoquartzitic beds.

Spec. No.	102.2	019.5(3)	019.4	019.5(2)
Quartz	49.8	45.0	40.9	50.9
Multigranular Quartz	6.4	3.8	1.1	3.7
Secondary Quartz	11.6	10.6	4.0	12.5
TOTAL QUARTZ	67.8	59.4	46.0	67.1
Matrix	19.8	23.8	40.0	24.0
Shale fragments	3.4	4.2	0.7	0.8
Kaolinite	1.2	1.6	-	0.3
Carbonaceous	5.8	7.4	11.8	2.6
Mica	2.0	1.4	0.4	0.8
Others	-	-	-	0.3
Voids	-	2.2	1.1	4.1
	d division	lamination	dark silty top, d-e divisions	Contorted fine sand above ripples of c division

greywacke but many specimens however do not have a typical greywacke texture and could be called 'coarse non-feldspathic clay-rich siltstones' (Ibid p.241).

The important point to be stressed is that there is a gradation from typical protoquartzite in the a, b and c divisions of the turbidite to the clay rich d and e divisions. In this gradation detrital quartz grain size and quantity decreases progressively and clay matrix becomes more abundant. There is also an increase in carbonaceous material towards the tops of the beds. This reflects increasing deposition of fine material in the later low velocity stages of the current and indicates that the majority of the fine grained material was kept in suspension during deposition of the a, b and c divisions. It thus follows that the turbidity currents were probably well sorted with regard to grain size particularly in the more distal areas.

REPLACEMENT OF PROTOQUARTZITES BY SIDERITE.

The protoquartzites discussed so far and modally analysed have been free of siderite as far as could be detected. Many beds are however replaced by siderite to a variable degree. Fig.2.1Fg shows siderite forming the cement to a sandstone and Fig.2.1Fh gives modal analyses of various beds indicating the variation in proportions of siderite and quartz found in the beds.

Siderite probably crystallized early in the diagenesis of the sediment, and it seems that it formed first between detrital sand grains where it probably replaced original matrix material (see Dapples 1967 p.104). This is borne out by the frequent absence of any other matrix material in the siderite cemented beds. Siderite also replaced quartz to such an extent that relict quartz grains are now seen 'floating' in siderite (Fig.2.1Fi). The ultimate result of this replacement process is a bed of siderite. Such beds occur and are thought to represent replacements of d-e type turbidites. Some siderite beds contain quartz grains at their base still arranged in parallel laminae and the thicknesses of the siderite beds (2-8cm. usual) are similar to those of d-e turbidites. Further support for this theory is given by beds of the type illustrated in Figs.2.1Fj-k. In Fig.2.1Fj the silty top to a turbidite is completely replaced by siderite which forms a continuous bed above the sandy portion of the turbidite. Fig.2.1Fk shows a bed in which nodules of siderite are developed in the silty top of the bed which is elsewhere preserved in the usual manner. The nodules apparently formed before compaction of the muddy top of the turbidite took place, thus indicating that siderite deposition took place early in diagenesis soon after deposition of the bed.

Holdsworth (thesis 1963 p.242) confirmed by X-ray diffraction that siderite was the only carbonate phase present in the beds, and recognised that the siderite crystallized early in the diagenesis of the sediment since uncrushed plant fragments occur within the siderite beds (Ibid p.244).

Fig.2.1 Fh

Modes of Siderite Cemented and Replaced Beds.

Spec. No.	351.4ab	351.3	002.4
Quartz	61.6	65.6	21.8
Siderite	36.8	29.4	76.8
Other	1.6	-	1.4
Holes	-	5.0	-

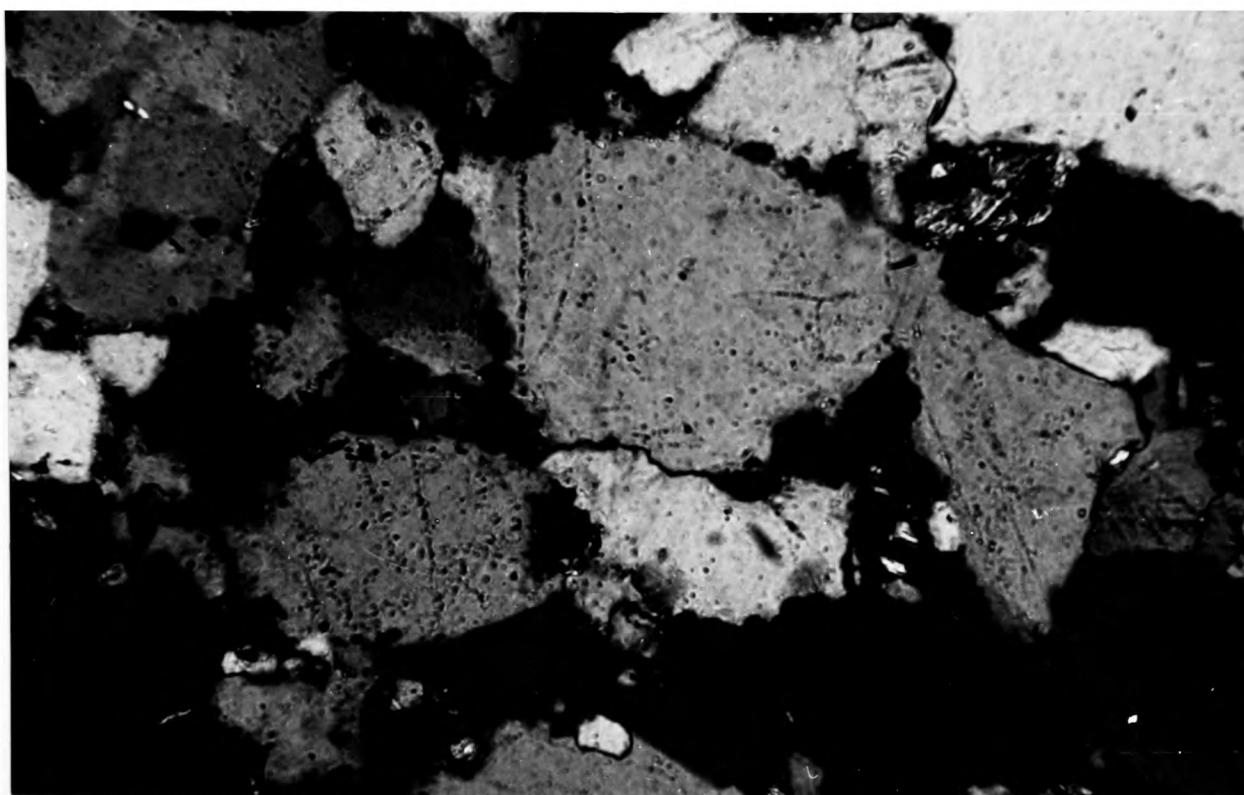


Fig. 2.1Ff

100 μ

Protoquartzite showing sutured contacts between quartz grains, due to secondary silica cementation.
Slide 009.5. Crossed polars. Unit C. Loc.009.

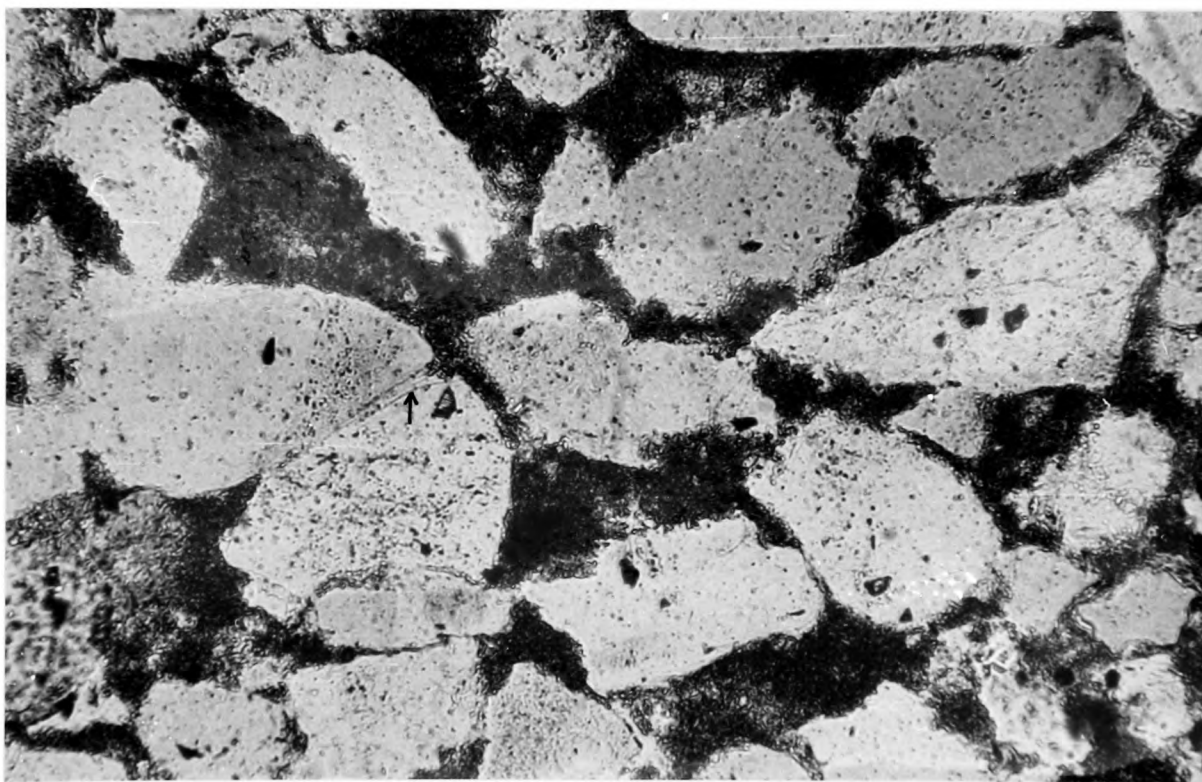
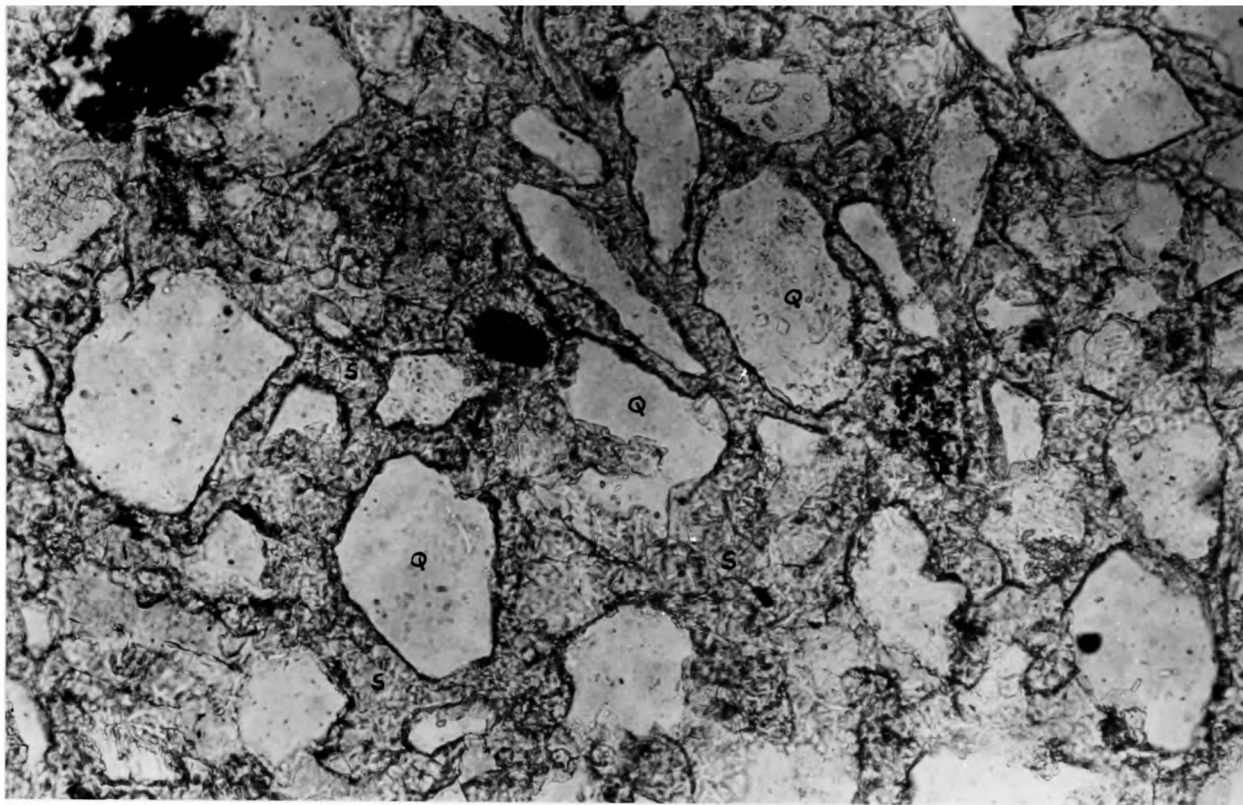


Fig. 2.1Fg

100 μ

Siderite (dark) cementing and replacing quartz grains, secondary quartz cementing grains indicated by arrow.
Slide 351.3. Base of Unit C. Loc.351.



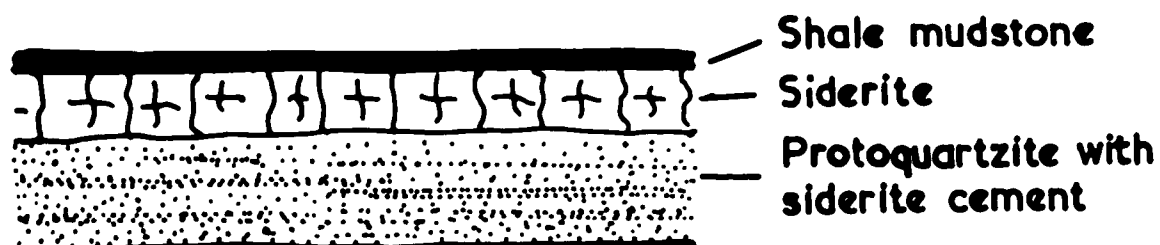
50 μ

Fig. 2.1Fi

Protoquartzite bed replaced by siderite so that quartz relics (Q) are left isolated in a matrix of siderite (S).

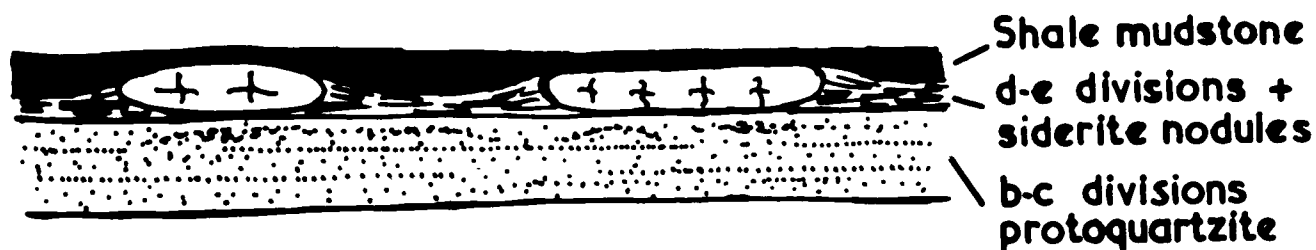
Slide 301.1. Unit C. Loc.301.

Fig 2.1Fj.



| 10cm.

Fig 2.1 Fk.



| 10cm.

Sketches of two protoquartzite beds to illustrate the replacement of the d-e divisions by either a continuous bed or isolated nodules of siderite. Both examples from Unit B. Loc.204.

2.1G

ORIGIN OF MATERIAL IN PROTOQUARTZITES.COARSE GRAINED MATERIAL.

Sections of the coarser grained, and sometimes pebbly proximal protoquartzites show a variety of grains derived from several different types of source rock. A brief examination of these was made in an attempt to determine the rock types present in the source area of the protoquartzites.

The vast majority of grains consist of quartz which contains few inclusions. Undulose extinction is fairly common but many grains show perfectly normal extinction. Very few grains show any regular arrangement of inclusions. Polygranular grains are of very frequent occurrence (over 50% detrital quartz) in the coarser grained specimens but in the finer grained rocks are less abundant, as would be expected. Polygranular quartz grains may show their constituent parts (Fig. 2.1Ga) or the boundaries may be straight. The sutured contacts suggest derivation from quartz rich metamorphic rocks or quartz veins whilst the straight contacts are more suggestive of an igneous origin and could indicate grains derived from granites or similar rocks. The scarcity of grains with regular mineral inclusions may indicate that metamorphic rocks are not the usual source material of the protoquartzites.

Grains of igneous origin.

Grains that are certainly of igneous origin are very scarce and never form a significant proportion of the rock. The most common grains seen that are thought to be igneous in origin are a brownish colour in ordinary light and consist of fine grained quartz with patches of amorphous material which sometimes have straight edges suggesting that they were once crystalline, but no clue remains as to their original nature. These fragments may possibly be weathered fragments of rhyolites but no trace of any original K-feldspar is seen.

A grain of definite igneous origin, and probably volcanic, is shown in Fig. 2.1Gb . Laths of twinned plagioclase are present in a highly quartz rich matrix, part of which may well be secondary, since it is known that secondary silification commonly affects the proto-quartzites (see section 2.1F).

A few grains were found with a marked igneous texture (Fig. 2.1Gc), which appeared to consist of feldspar laths in a matrix, but which under crossed polars were found to consist of large crystals of quartz which cut across the igneous texture of the grain. The other material present consists of patches of chlorite and iron hydroxides. The fragments appear to be

silicified lava, and were possibly originally basaltic judging from the general texture. Whether the silification took place before or after deposition it is not possible to say. Occasional large quartz grains contain euhedral apatite crystals (Fig. 2.1Gd) and closely resemble the apatites frequently seen in granitic rocks.

Grains of Sedimentary Origin.

Grains clearly derived from sediments are infrequent, apart from fragments of dark shale and mudstone which were probably derived from erosion of the sea floor over which the currents flowed. The majority of the quartz grains could equally well have come directly from igneous or sedimentary rocks and it is impossible to distinguish them. The rounding of the quartz grains present varies considerably - the majority of grains are angular but a few are well rounded and may be recycled sedimentary grains or merely derived from a more distant part of the source area and thus have travelled further to the depositional area.

Some grains are clearly derived from previously existing mudstones (Fig. 2.1Ge) but these are rare. Grains of banded chert (Fig. 2.1Gd) are clearly detrital in nature, and could be of sedimentary origin, but equally

could have been a secondary siliceous filling of cavities in almost any type of rock.

Grains of Metamorphic Origin.

Some polygranular quartz grains show regular elongation of strained quartz crystals, and appear to be derived from quartzose schists, such grains are rare. A very few grains were seen containing aligned micas (Fig. 2.1Gf) and such fragments are derived from quartz-mica schists.

Some grains (Figs. 2.1Gg-h) were derived from low grade metamorphic rocks of a slaty type. They contain chlorite and fine grained quartz and occasional small flakes of mica which have all been recrystallied.

Other grains which may be of metamorphic origin consist of quartz with enclosed vermicules of chlorite (Fig. 2.1Gi). The vermicules were clearly in place before deposition of the grain as they are truncated at the grain boundaries. Similar quartz containing vermicular chlorite has been observed by the author in low grade (chlorite) schists and also in material from quartz veins traversing low grade metamorphic rocks (e.g. Snowdon area, North Wales).

Negative Features.

K-feldspar is absent from even the most proximal protoquartzites. Numerous sections were stained with

potassium cobalti-nitrate but no K-feldspar could be detected. Holdsworth (thesis 1963) also considered the protoquartzites to be free of K-feldspar. The absence of K-feldspar could be due to the absence of this mineral in the source area, but is more probably due to its removal during the weathering and transport of material. The numerous plant fragments in the protoquartzites indicate derivation of the material from a well vegetated area. Acid conditions prevailing in such an area especially under the warm conditions associated with the Carboniferous would have made weathering of K-feldspar rapid. Any K-feldspar that reached the depositional area could also have been altered to kaolinite by acid conditions prevailing below the sediment surface. In the protoquartzites patches of vermicular kaolinite have been observed (Holdsworth thesis 1963) that possibly represent altered K-feldspar. Grains that suggested original rhyolitic fragments (p.203) have kaolinite included in them and this may also represent original K-feldspar.

FINE GRAINED MATERIAL.

The green silty tops of several of the turbidites were examined by X-ray diffraction. The principal clay minerals present are kaolinite and chlorite and they are

associated with well crystallized detrital muscovite mica. A marked tail on the low angle side of the 10\AA mica peak represents mica or 'illite' being altered to mixed layer mica-montmorillonite. These minerals are the same as those found in the shale-mudstones (Sect. 1.4) but here they are better crystallized probably reflecting the larger grain size of the silty turbidite top and the fact that it was deposited quickly. Also the greater admixture of fine carbonaceous material in the shale-mudstones tends to mask the clay mineral reflections.

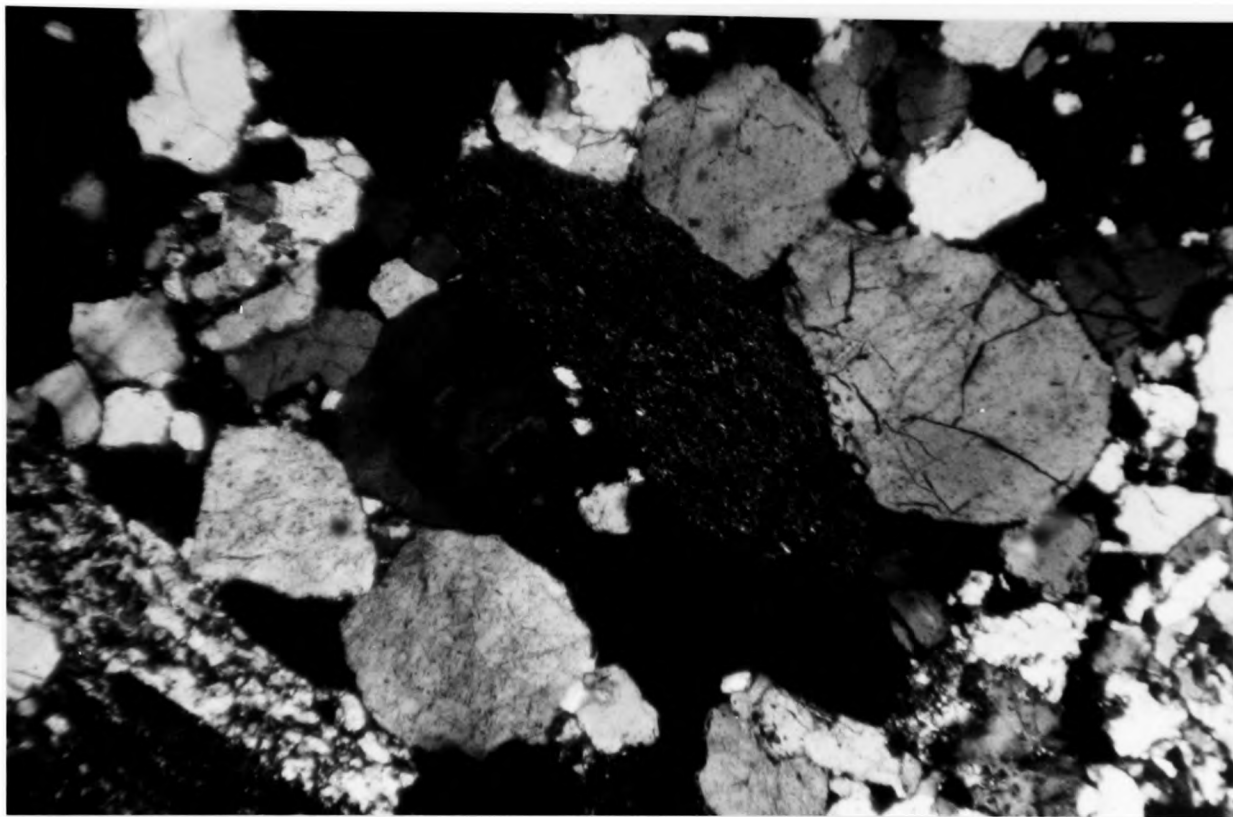
The kaolinite is ~~is~~probably partly detrital and partly diagenetic in origin, and represents the weathering product of feldspar. The chlorite could be derived from the weathering of ferromagnesian minerals in the source area derived from igneous rocks, but could equally be derived from chlorite grade metamorphic rocks. Both igneous and meta-sedimentary fragments containing chlorite have been seen in thin section and so the chlorite is probably derived from both sources. It is of interest to note that the present soil of North Wales (Ball 1966) is very rich in chlorite and the predominant rock types are of chlorite grade and include both igneous and sedimentary types.

The mica present also has several possible sources

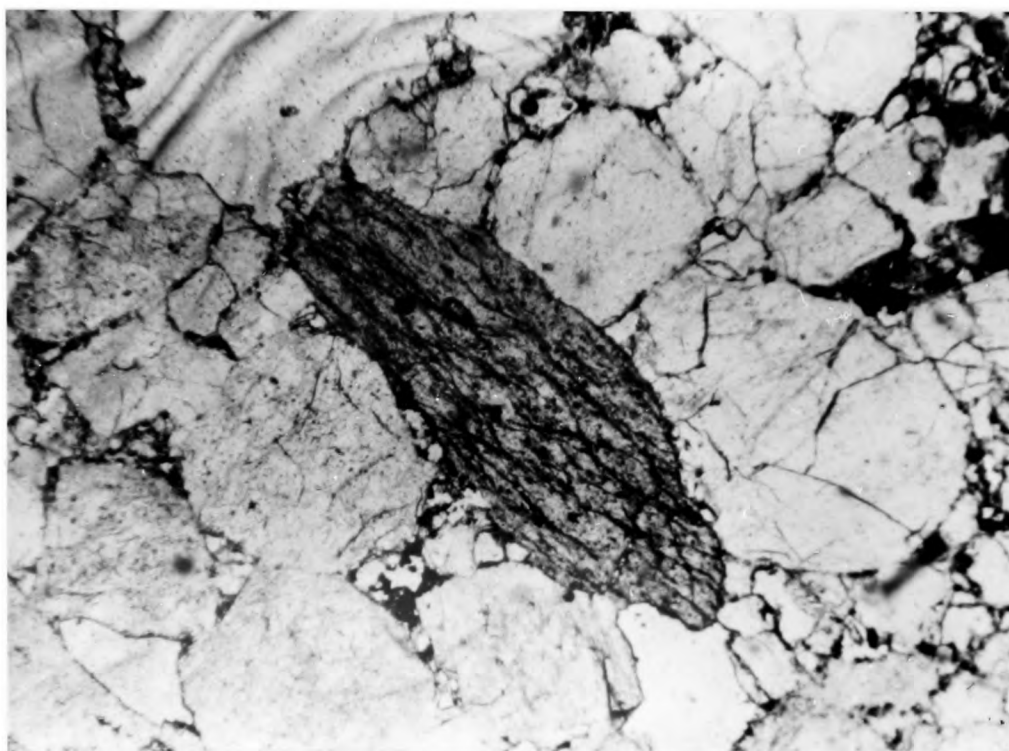
all of which probably contributed part of the material. The 2M high temperature polymorph of Yoder and Eugster (1955) has been recognised and thus some of the material is derived from igneous rocks such as granites or from metamorphic mica bearing rocks. The mica could also have been recycled via a sediment. Some micaceous material that is poorly crystalline may be diagenetic.

The bulk of the chlorite and kaolinite is considered to be detrital in origin and derived from ~~the~~ weathering under acid conditions of feldspar bearing rocks such as rhyolites or granites together with chlorite grade metamorphic rocks of igneous and sedimentary origin. An area similar to the Lower Palaeozoic outcrop of Wales would appear to constitute a suitable area.

Fig. 2.1Gh



Crossed polars.



500 μ

Greenschist fragment (fine grained quartz and chlorite)
in protoquartzite bed. Slide 073.2E. Unit B. Loc.073.

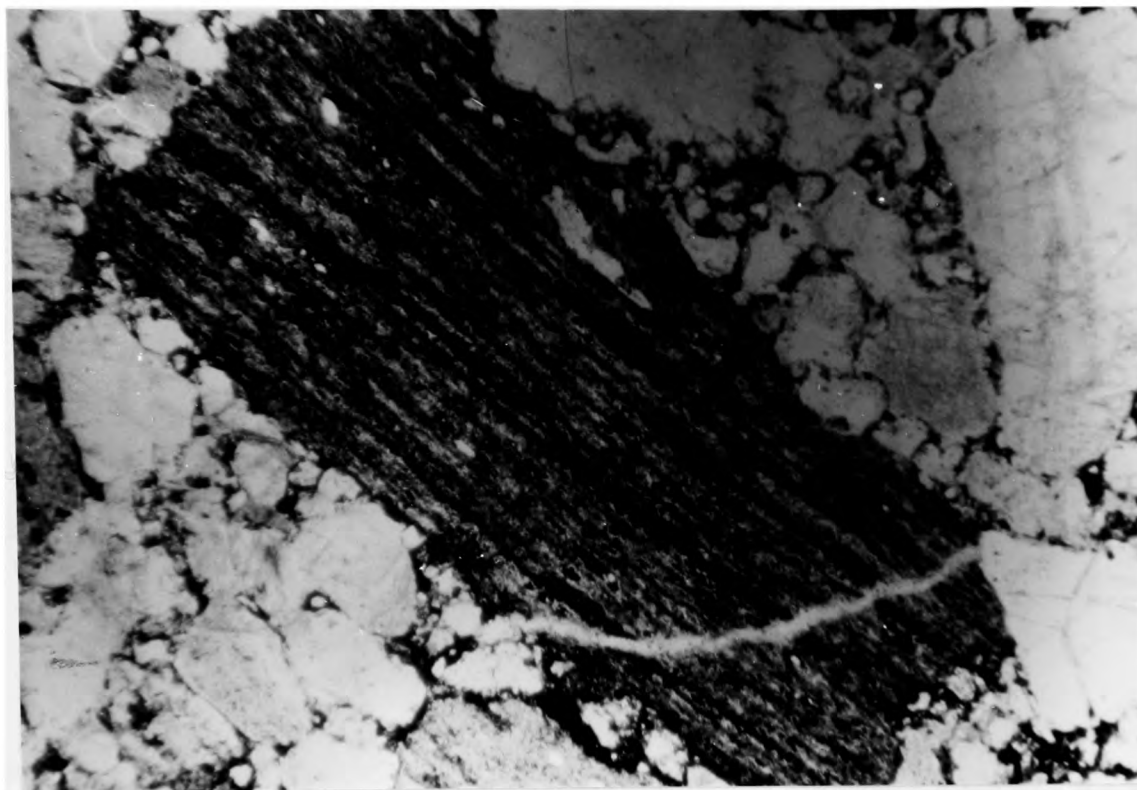


Fig. 2.1Gg

Detrital grain of slate traversed by a quartz vein.
Slide 073.2. Unit B. Loc.073.

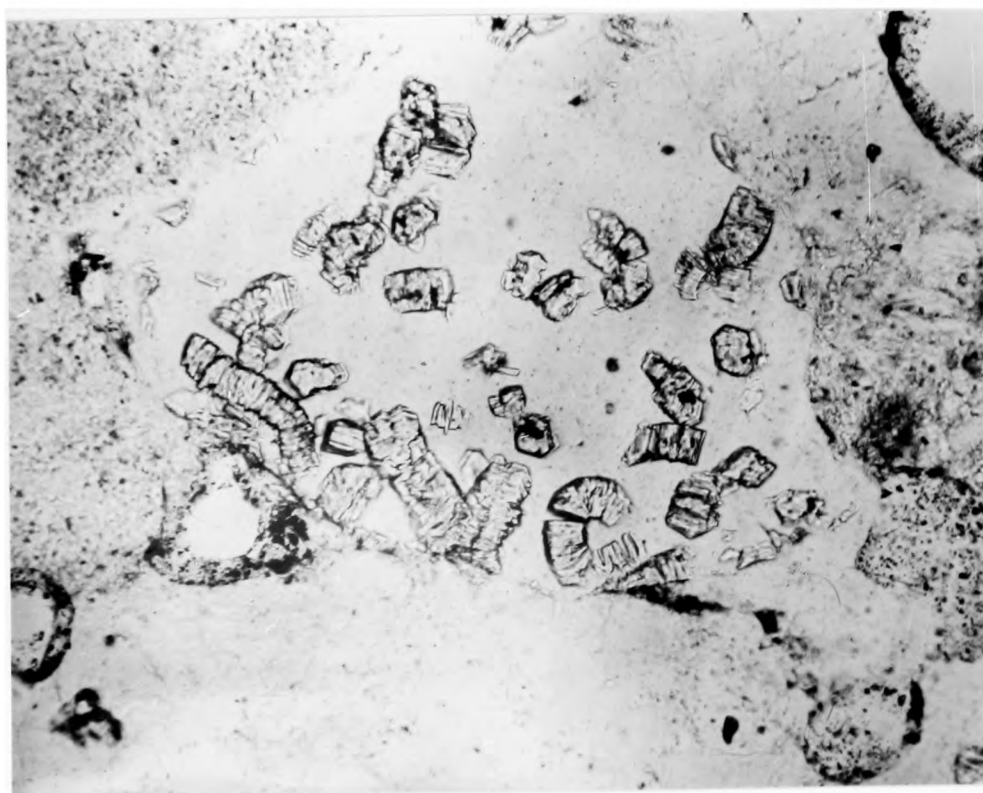


Fig. 2.1Gi

Chlorite vermicules in a quartz grain, vermicules are truncated at the grain boundary. Slide 200P, Protoquartzites H zone. Lumpool Bridge, Staffs.

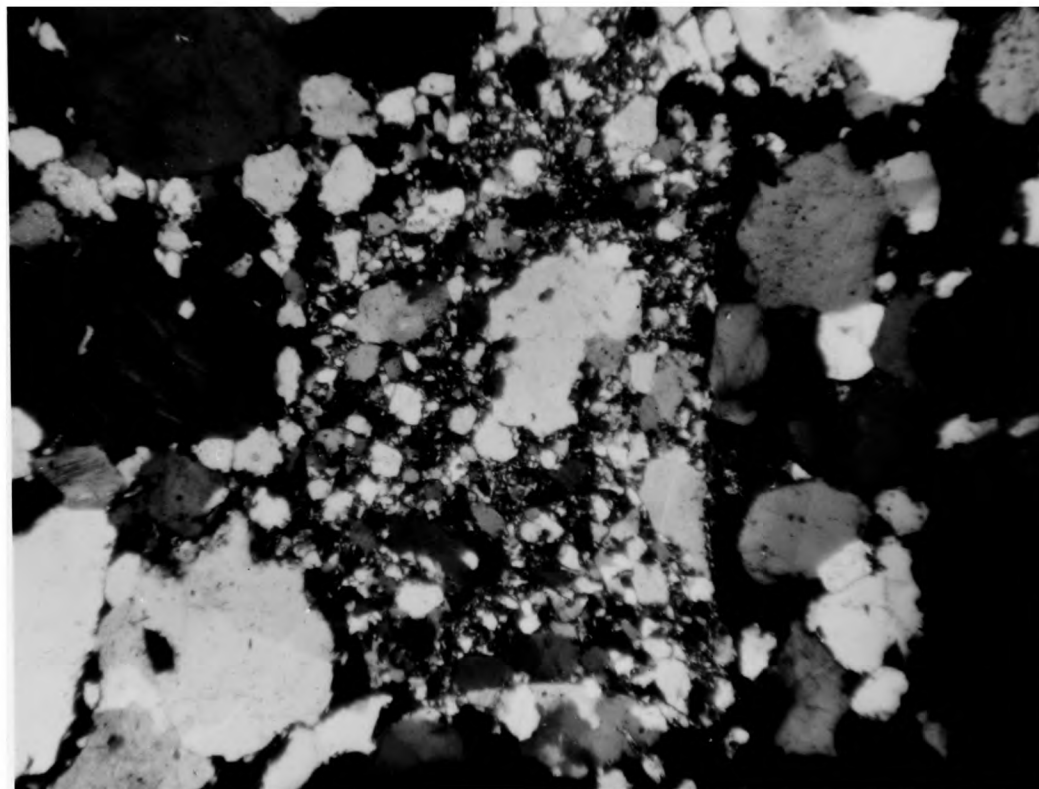


Fig. 2.1Ge

500 μ

Detrital grain of sandstone included in a protoquartzite bed.
Slide 073.2c. Crossed polars. Unit B. Loc.073.

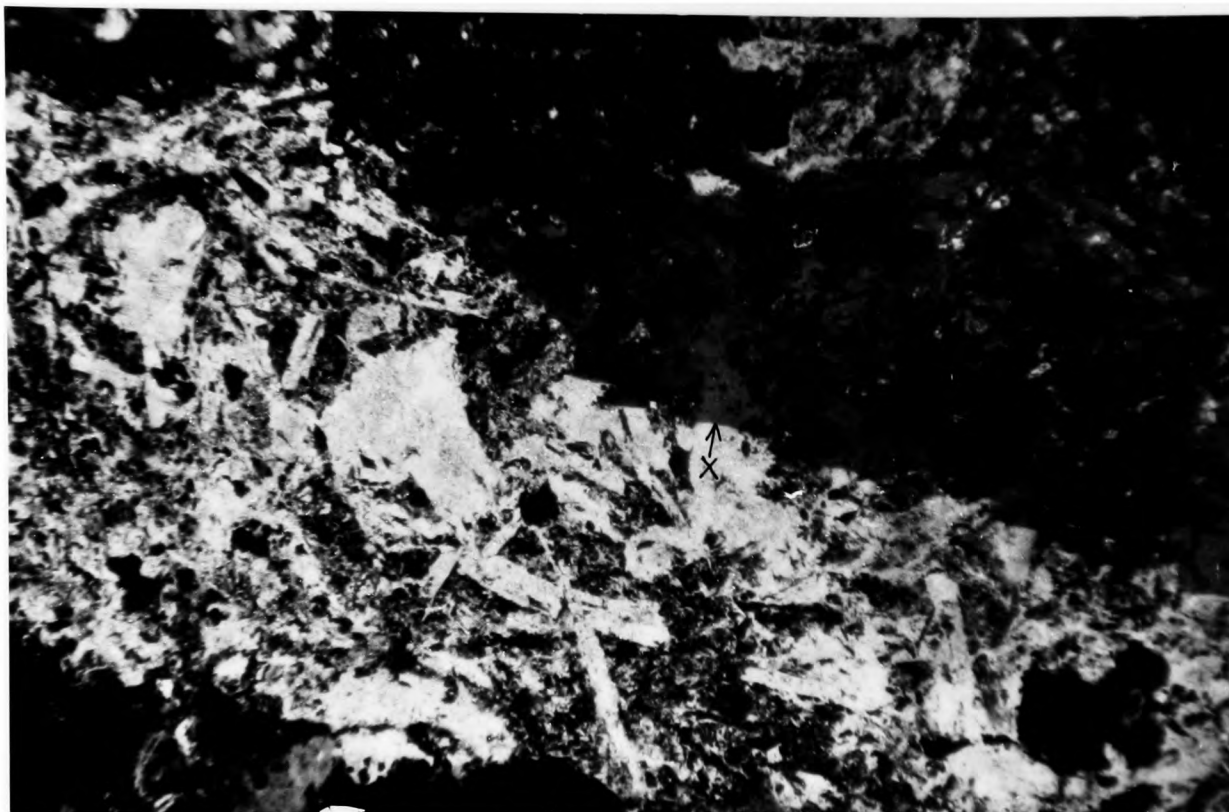


Fig. 2.1Gf

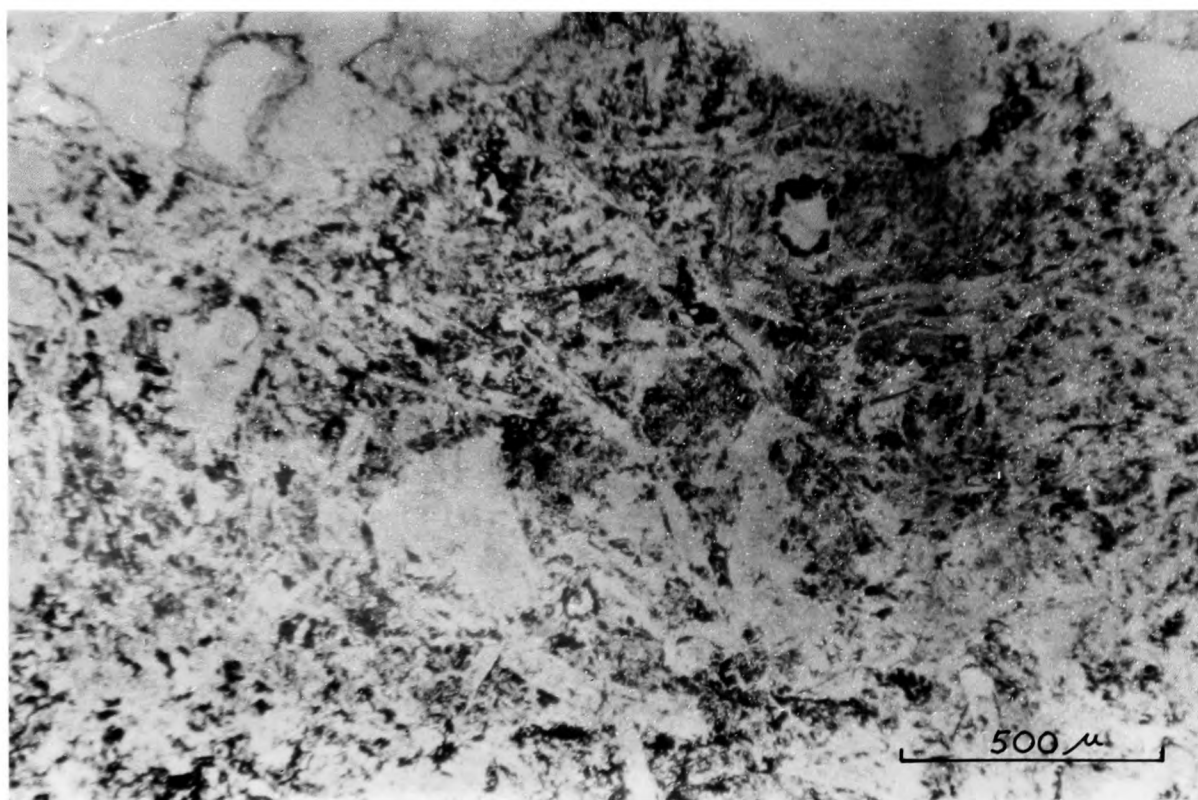
500 μ

Grain of quartz-mica schist included in protoquartzite bed.
Slide 073.2c. Crossed polars. Unit B. Loc.073.

Fig. 2.1Gc



Crossed polars.



Grain with igneous texture but silicified and consisting of quartz with a little chlorite. Lath shapes were originally feldspar but are now entirely quartz, and quartz grain boundaries cut laths as at x.

Slide 200P. Protoquartzites, H. zone. Lumpool Bridge, Staffs.

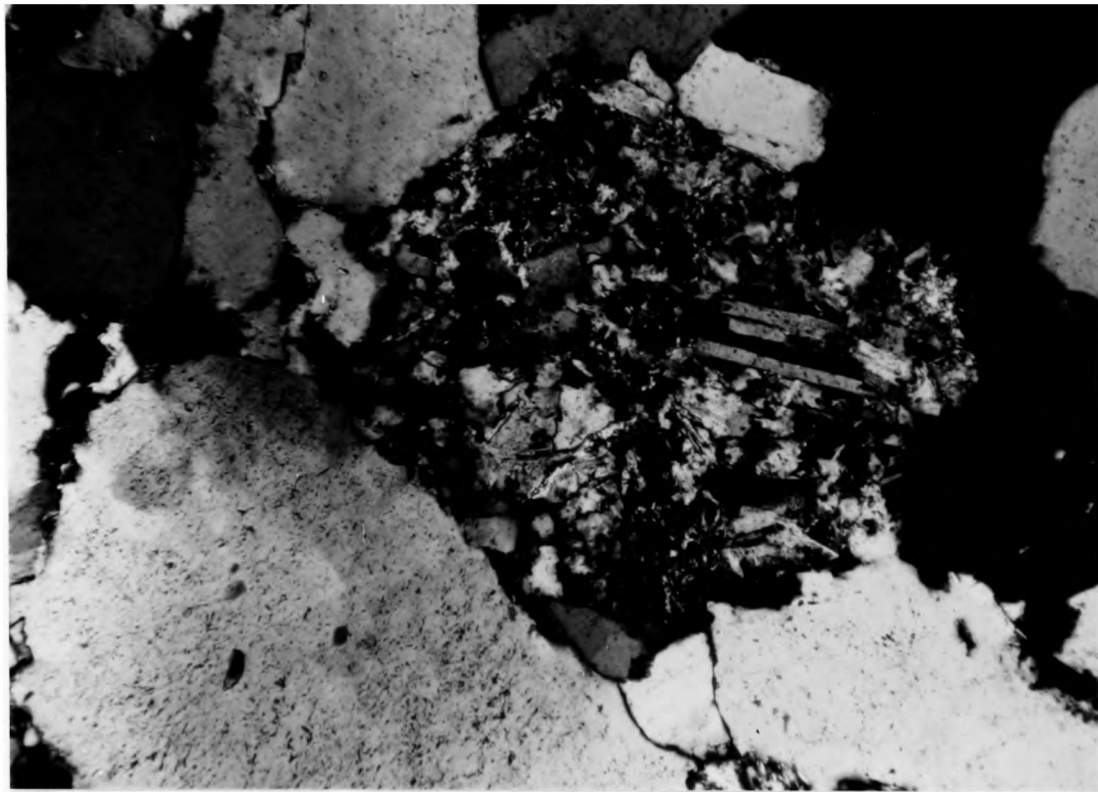


Fig. 2.1Gb

100 μ

Grain of igneous origin in protoquartzite bed, feldspar is albite. Slide 010.1. Crossed polars. Unit C. Loc.010.

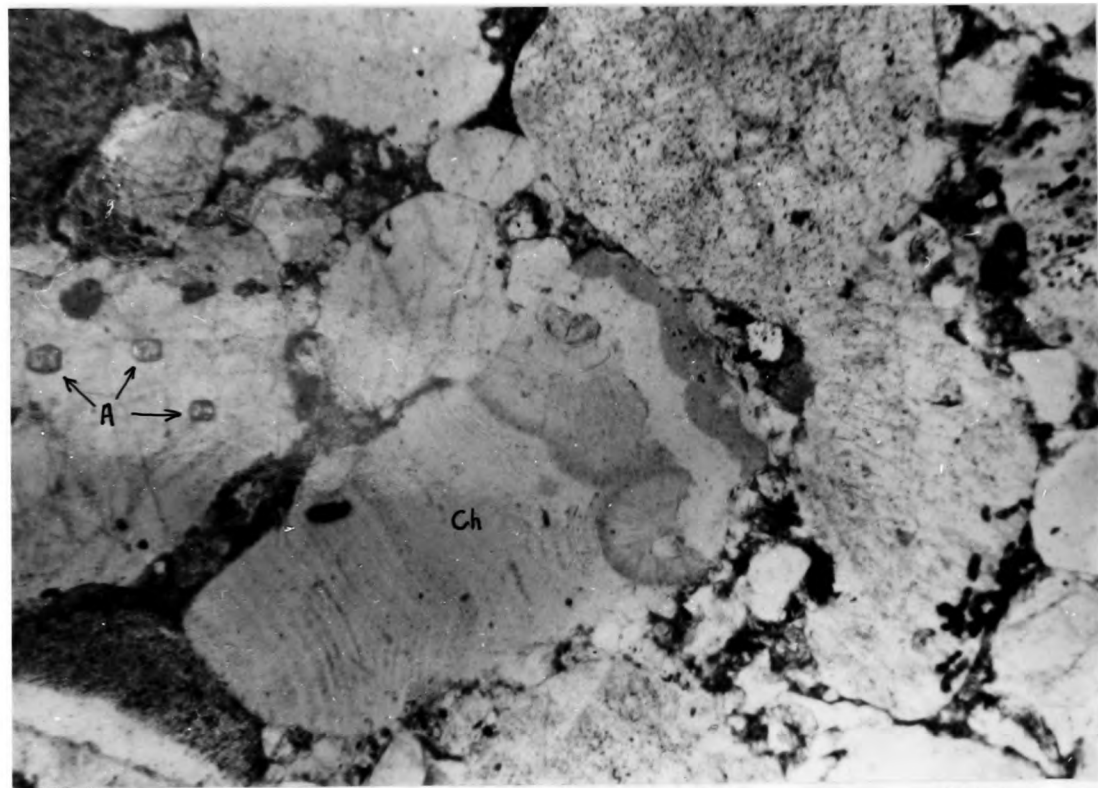


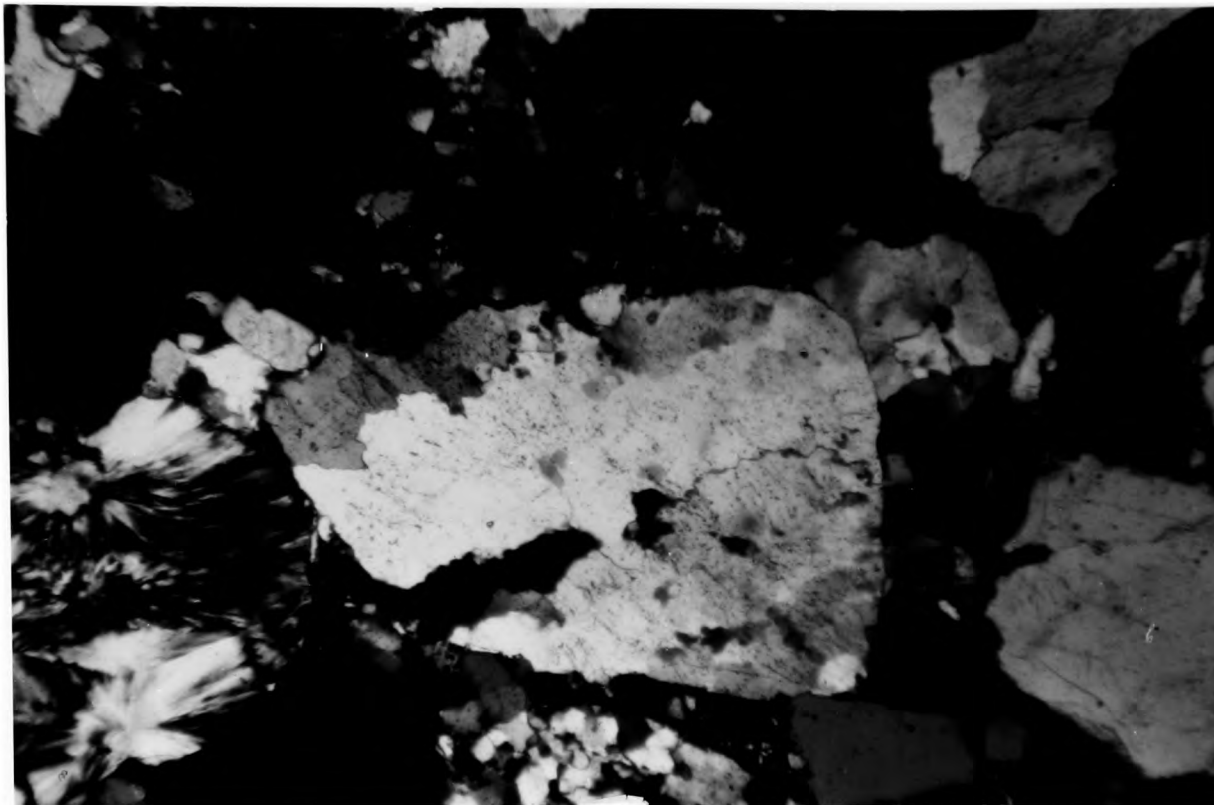
Fig. 2.1Gd

500 μ

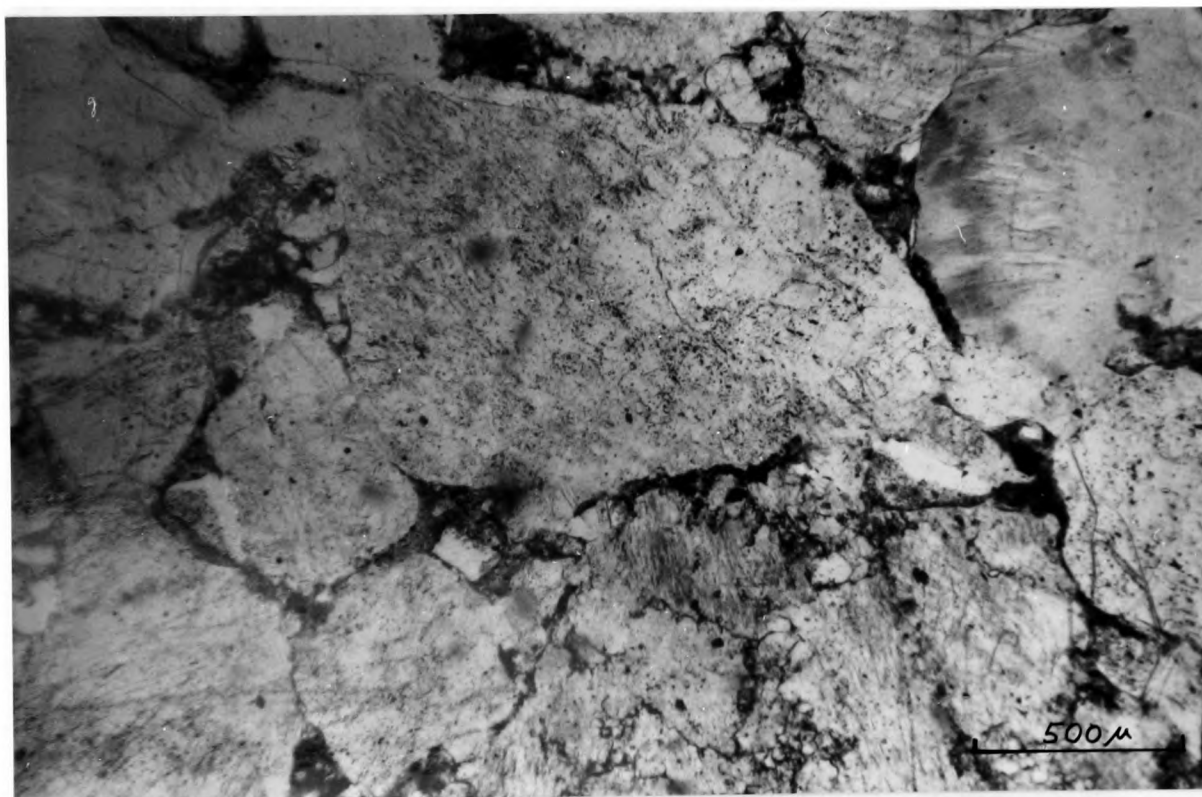
Detrital banded chert grain (Ch) and quartz containing euhedral apatite crystals (A) in protoquartzite.

Slide 072.1A. Unit B. Loc.072.

Fig. 2.1Ga



Crossed polars.



Sutured contacts in polygranular quartz grain. Chert grain at left. Slide 073.2. Unit B. Loc.073.

2.1H

ORIGIN OF THE PROTOQUARTZITES.

The general features of the protoquartzites indicate that they are probably turbidites. The sequence of internal structures and the assemblage of sole markings are identical with those described by many authors (Dzulynski and Walton 1965 for bibliography) from turbidite successions.

In the protoquartzites tool markings are of much more frequent occurrence than scour markings and it is likely that the formation of bottom structures was controlled by the nature of the turbidity current. The protoquartzites are generally similar to mineralogical content consisting mainly of quartz grains with little admixture of feldspar or rock fragments (see section 2.1F). Plant fragments are of frequent occurrence and often form dark laminations along which the protoquartzite splits.

The types of grading observed are typical of those found in the turbidites. The distal beds, where grading was most frequently seen were probably deposited by mature flows with well developed horizontal and vertical grading which caused the clay and silty material to be held in suspension during the deposition of the lower divisions, and deposited in the final stage of flow

as the d-e divisions.

The proximal protoquartzites are interpreted as being deposited by immature turbidity currents which had not achieved any significant measure of vertical or horizontal grading. Dzulynski et al. (1959) proposed the term 'fluxoturbidite' for such deposits which they considered to have been deposited from an immature turbidity current flowing as a watery mass. They envisaged the sequence Slump - Immature turbidite - Mature turbidite in which a fluxoturbidite was the deposit formed in the transition from slump to turbidity current. The term fluxoturbidite has been used by many authors for a variety of beds which do not all fit the original criteria of Dzulynski et al. (1959). The most significant features about these beds are their lack of grading, except in the top few centimetres, non-development of the internal structures described by Bouma (1962), frequent inclusion of coarse material and poor sorting, and indistinct bottom structures with load casts most abundant. In the present area it can be shown that in the downcurrent direction the beds thin markedly, and develop internal and external structures typical of turbidites. Thus they are considered to represent the deposits of immature turbidity currents which as they flowed out into the basin became better

sorted and deposited the turbidites found there.

The features displayed by the proximal proto-quartzites thus fit the features assigned to fluxoturbidites by Dzulynski et al. (1959). Many other authors have described fluxoturbidite sequences (Walker 1967 p.39 for summary), and there is a general measure of agreement about the structures present. Walker (1967 p.41) prefers the term 'Proximal Turbidites' to Fluxoturbidites since little is known about their mode of transport. The 15 fluxoturbidite features listed by Walker (1967) are shown in (Fig. 2.1Ha), together with the observations of Dzulynski et al. (1959), Walker (1967) and those from the present proximal protoquartzites. The features observed in the proximal protoquartzites correspond well with those of Walker (1967) and Dzulynski et al. (1959) and the interpretation of the beds as proximal turbidites appears to be justified.

As has already been mentioned, base cut-out sequences are of frequent occurrence in the distal protoquartzites of the basin area and these are taken to represent the deposits of waning turbidity currents which had deposited the bulk of their load and were only carrying the finer grained material at a lower velocity. Walker (1965, 1967) has interpreted the internal sequence of structures in terms of the slowing down of the turbidity current, and considers divisions a and b

Comparison of the features of the proximal protoquartzites of north Staffordshire with those observed by Walker and Dzulynski et al. from other areas.

Fluxoturbidite features from Waller 1967	Dz.	W.	proximal protoquartzites north Staffordshire.	
1. Beds composed of unusually coarse sediment.	p	p	p	Pebbles to 2cm.
2. Beds thicker than usual.	p	p	p	Maximum bed thickness 1.5m. Usually 10-50cm.
3. Grading poor or absent.	p	p	p	Grading usually only apparent in top few cms., if present at all.
4. Multiple beds, repeated grading common.	p	p	p	Multiple beds frequent, repeated grading present but not common due to 3 above.
5. Mudstone partings thin or absent.	p	p	p	Mudstone very rare. Pebbly-sand or silt often present between beds.
6. Beds irregular, not parallel sided.	p	p	p	More irregular than distal beds on bases often due to loading.
7. Beds interbedded with turbidites.	p	p	p	Interbedded with 'Thin bedded proximal developments'.
8. Sole marks scarce.	p		p	Load structures more common ; scour and tool marks rare.
9. Sandstones cleaner than usual.	p		-	Not applicable as sandy parts of beds are usually protoquartzitic in both proximal and distal areas.
10. Mud flakes included in sandstones.		p	p	True of both proximal and distal developments.
11. Large scale cross-stratification present.	p		p	Only one example seen; planar cross section 20cm. thick.
12. Scouring and channelling common.		p		Not observed.
13. Sandstones have sharp tops as well as bases.		p	p	Beds with sharp tops are usually rippled at their tops -
14. Sandstone tops have lingoid ripples.		p	p	this feature also applies to distal beds.
15. Beds show indications of 'slumping' or 'sliding'.	p		p?	Small slump folds apparent in beds interbedded with 'Proximal' beds i.e. Thin bedded proximal developments.

p = presence of feature.

Dz = Dzulynski et al. (1959 p.1114).

W = Walker (1967 Tables 2 & 3).

to be formed in the upper flow regime (Simons et al. 1961) and divisions c to e in the lower flow regime. Division a contains no preserved structures since the bed form was not in equilibrium with the current and was destroyed as the current slowed down. The lower laminations represent the stable bed form in the lower part of the upper flow regime. The general absence of the dune phase (Simons et al. 1961) in turbidites is explained by Walker (1965) in terms of insufficient depth of flow in the current, or lack of time for the dunes to form before the current slowed down to rippling velocity. Walton (1967) considers the possibility that dunes may be suppressed for certain modal grain sizes, and notes that at the stage when dunes would be expected the available material is often fine-grained sand and this might be lifted into suspension instead of moving as a bed load and forming dunes. There is also a suggestion that the range of grain sizes present, especially the presence of fine grained material, may inhibit dune formation. Simons et al. (1963) showed that the dune phase was not developed when the current contained a high proportion of fine material. This was due to the development of a cohesive sediment surface in the preceeding ripple stage. In a turbidite, this surface would not be present as the current is slowing down rather than

accelerating and could not affect dune configuration. In their 2 foot wide flume Simons et al. (1963 p.632) were able to watch the progressive reduction in the amplitude of dunes on increase of the proportion of bentonite clay in the water. This case does have more relevance to a turbidity current, and lends support to the contention that suspended clay material inhibits development of the dune phase in turbidites. Clay material was certainly abundant in the turbidity currents responsible for the deposition of the proto-quartzites and it seems that most of the clay was held in suspension during the deposition of the a, b and c divisions since it is always concentrated at the top of the deposit.

Hubert (1966 p.685) has recognised the dune phase in graded limestones (Whitehouse Formation, Ordovician, Girvan, Scotland) which have similar internal and external structures to turbidites but which Hubert points out (Ibid p.689) could be deposited by waning bottom currents.

The ripple interval within the deposited turbidites can be observed to come into being gradually out of the laminated interval. When the b division is split along its laminations it is frequently noticed that the laminations are not of even thickness and there seems

to be a gradation into ripples of Type 3 of Walker (1963). Thus the author considers there is an upward gradation from laminations to ripples. It is already established that the ripple interval dies out gradually upwards (Walker 1965 p.15) with decreasing amplitude of the ripples until parallel lamination again appears (d division). This illustrates the transitory nature of the structures developed in turbidites during deposition. It must always be stressed that the structures described are formed under conditions of net deposition and from a decelerating sediment laden current containing a variety of grain sizes. These conditions are unlike those under which many experiments on bed forms have been performed, where considerable time has been allowed to elapse for stabilisation of bed form under constant flow conditions and using selected sediment grain sizes (Simons et al. 1961, 1963). It seems probable that the conditions are so different that the degree of correlation already achieved between structures in turbidites and bed forms obtained in flumes is quite remarkable.

Simons et al. (1963) have studied the effects of the presence of suspended fine grained material within a current on the bed form. The presence of clay material in concentrations up to 10% increases very considerably the viscosity of the liquid and profoundly

affects the stable bed form. The fine material was found to have a tendency to stabilise bed forms, and ripples developed rounded rather than angular crests. One significant observation made is that the fine grained material inhibits the development of dunes when the velocity is increased. Care must be taken in comparing this situation where the current is increasing in velocity with that in a turbidite current where velocity is decreasing during deposition, but it seems likely that the presence of fine clay material in the flow profoundly alters the bed form.

In the present work it was found that the ripple division was frequently absent giving the division sequences ab - de; b - de. Reasons for the absence of the c division are likely to be of a similar kind to those proposed for the absence of the dune phase. It is also likely that the rate of deceleration of the current has a marked effect on the structures produced (Walton 1967 p.311). If a turbidity current enters a region of gentle downhill slope then gravity is acting with the current and deceleration would be expected to be slow. In this case the current would be flowing for appreciable periods within specific velocity ranges, and thus there would be time for structures such as ripples to form. A second case can be imagined in

which a turbidity current enters an area of gentle uphill slope and is acting against gravity in having to flow uphill. In this second case deceleration would be much more rapid and the current would be flowing for a shorter period of time within a specific velocity range and the stable bed form may not have time to develop.

The first case described could be expected to give turbidites with an extensive depositional area in a downcurrent direction, and well developed internal structures stable over long distances. The second case would give turbidites reducing quickly in thickness in the down current direction and less well developed structures. There would possibly be a tendency for the flow to fan out more in a down current direction, both cases are shown diagrammatically in Fig. 2.1Hb. It is possible that the turbidity currents in the north Staffordshire area had to flow uphill on approaching the old Viséan reef line and thus the frequent lack of ripples within beds could be due to fast decelerating currents having to flow slightly uphill.

In case b, (Fig. 2.1 Hb) the bulk of the coarse material carried by the current would probably be deposited near the break in slope thus giving beds which thin more rapidly in a down current direction than those

Fig 2.1Hb
Basin slope

a

Deposited bed

Area of deposition



Basin slope

b

Deposited bed

Area of deposition

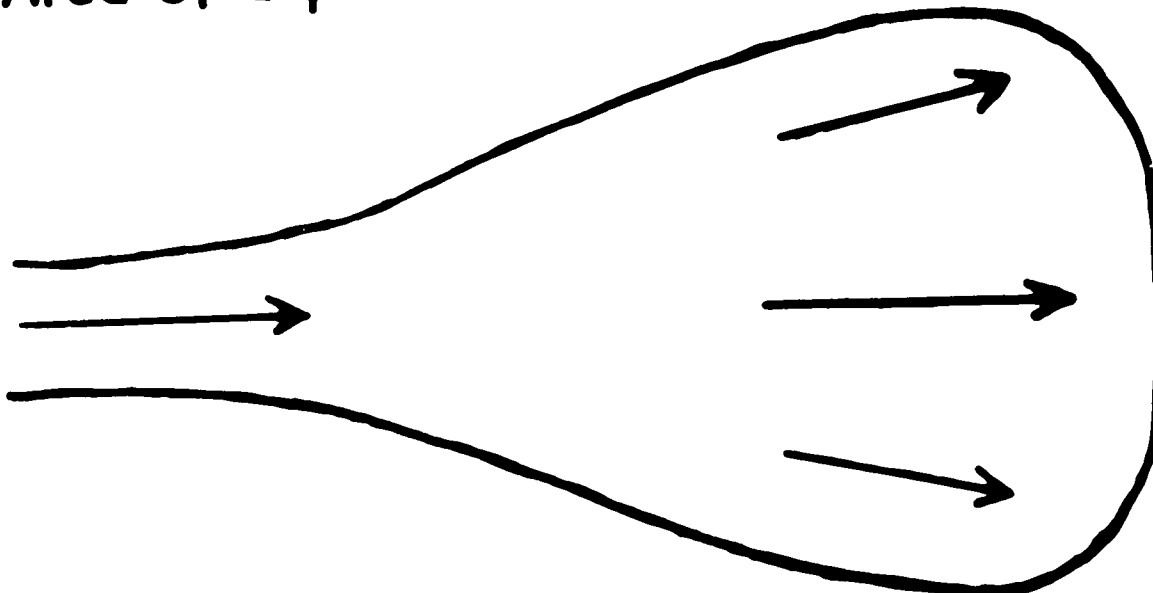


Diagram to illustrate the possible effect of basin slope on the thickness and depositional area of a turbidite.

deposited on a flat or downward sloping area. Also it is possible that the proximal deposits on the downward slope would be thinner than those at the start of an upward slope since more material could be kept in suspension as the current flowed downhill.

Deceleration of the current to velocities below that required for ripples to be formed is presumed to give rise to the d and e divisions of the turbidite. The exact mechanism of the formation of laminae is still very poorly known and the experimental evidence available is insufficient at present. Kuenen (1965) has suggested that similar grains tend to congregate in patches on a bed surface and give rise to lamination when covered with patches of other grains. This does not explain the lamination seen in the d division of turbidites where laminae are often continuous for several metres. Various authors have suggested that lamination is the result of bottom traction, (Dzulynski and Sanders 1962 p.87), but Walker (1965 p.19) considers this unlikely "because the ripples in the previous division stopped moving when traction ceased". It has frequently been observed that the ripples of the c division die out gradually upwards without any discontinuity until parallel lamination is attained, thus it would seem

from field evidence that there is a continuous gradation from ripple lamination to parallel lamination and that the processes of lamination formation are similar in both.

Several authors favour inhomogenities in the current as the reason for lamination formation. Changes in the velocity of the current at one point due to eddies, or intermittent supply of sediment from the current, (Bouma 1962 p.98).

Walker (1965 p.19) considers that the formation of the d division lamination is due to settling of grains through a laminar boundary layer, and states that:- "Graded and sharp based laminations could (therefore) develop by a process of intermittent supply of mixed sediment to the top of the laminar boundary layer." This is difficult to visualise in practice since it would require a current to start with a sediment free laminar boundary layer which was then suddenly supplied with sediment of mixed grain size from the upper turbulent part of the current. This would then have to settle through the layer without any other sediment being added to the top of the layer, (Fig. 2.1Hc). If the supply to the boundary layer was homogeneous and continuous there is no reason why lamination should develop except in the

form of a coarse base marking the start of the process and a fine top at the end of deposition (Fig. 2.1Hc). Thus Walker's explanation also involves intermittent supply of material from the turbidity current.

It is possible that the laminae are formed by the freezing of a traction carpet at the base of a laminar boundary layer. This could be due to reduction in velocity of the current or increased supply of material. During movement of the traction carpet, clay material could possibly be kept in suspension, but when the carpet reached a critical thickness or density, any slight variation in current or supply could cause it to freeze. This might result in the disruption of the laminar boundary layer for a while giving a deposit with clay until the boundary layer and/or traction carpet are re-established. (Fig. 2.1Hd).

At the time when the d division was being deposited, and also during the preceeding stages, the fine grained material was maintained in suspension, thus it would have profoundly influenced the sedimentation of the fine sand still present in the current. (Simons et al. 1963 p.G29). If the amount of fine material reaches a large enough concentration then it cannot be kept in suspension and is "deposited on, and in, the bed causing a cohesive stable boundary to form." (Ibid p.G29).

It is possible that this could happen during deposition from a turbidity current carrying a variety of grain sizes. The resulting deposit could be a graded lamina or there may be a sharp contact between sand and mud trapped on its surface. Subsequent passage of parts of the current containing less fine grained material could enable redeposition of sand to take place on the mud which could again become stabilised by the deposition of clay at the surface. This process would obviously be facilitated by the passage of variations in velocity or sediment concentration within the current.

This stabilisation of laminae due to deposition of suspended clay may also give an indication as to why ripples of Type 1 of Walker are found in turbidites and truncations of laminae are rare. A sandy lamina once formed could quickly be stabilised by deposition of clay material on its surface. The control of the deposition would effectively be the concentration of clay (and hence liquid viscosity) in the water flowing above the bed. The gradual reduction in ripple amplitude until plane bed is obtained being a reflection of continuous reduction in velocity of the current.

The thicknesses of the laminae are perhaps dependent on grain size and current velocity. The lamina is

possibly in motion on the bed due to current shear until conditions are attained when the lamina is stabilised by incorporation of clay in, or on, its upper surface (Fig. 2.1He). Thus a lamina may originate as a traction carpet.

Any remarks on lamination formation are obviously very tentative and the only way the problem of lamination can be solved is by experimental work with suitable materials under conditions of deposition.

The graded pelitic top to the beds is simply explained as the final settling of material from the tail of material left in suspension after the passage of the main body of the turbidity current. It represents the last deposit due to the turbidity current. In the present area the e divisions are green in colour and contrast well with the black shale-mudstone which is presumed to be the normal basin deposit. The black shale-mudstones are finer grained and richer in carbon than the e divisions, and contain clay minerals which are poorly crystallized when compared with those of the e division, (see section 2.2.).

ORIGIN OF PROTOQUARTZITE BEDS WITH RIPPLED TOPS.

It has already been noted (page 185) that a proportion of the protoquartzite beds have rippled tops against black shale-mudstone with the d and e divisions

Fig 2.1Hc

Intermittent supply

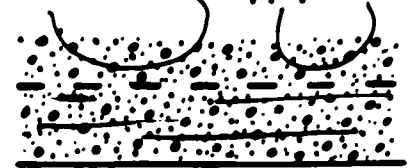


Laminar layer



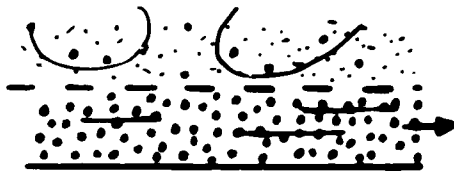
Graded lamination
formed

Continuous supply



No graded lamination
formed

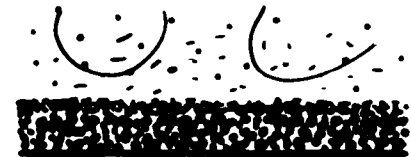
Fig 2.1Hd



Moving traction
carpet



Carpet freezes

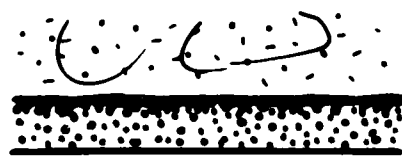


Different material deposited
from turbulent current

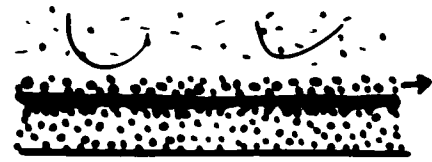
Fig 2.1He



Moving traction
carpet.



Traction carpet stabilised
by incorporation of
clay.



New traction carpet
established.

absent. It has also been observed that these ripples are of Type 1 (Walker 1963) and are due to reworking of the top of the sand layer.

On the turbidite hypothesis this is possible if a turbidity current is in a non-depositional phase (Walker 1965 p.15) and reworking of the top takes place. It is also possible that after the passage of the sediment laden current, a current of clear water followed in its wake reworking the top of the deposited sand and keeping in suspension the fine material that would have formed the d and e divisions.

Another possibility is that there was little fine grained material present in the current and that the 'tail' of the current was able to rework the sand into ripples until the bed became static, and no more material was deposited.

In both these cases the current directions indicated by the ripples should correspond roughly with the current directions indicated by the other structures displayed by the turbidites. (Walker (1967 p.38) considers that eddies following in the wake of a turbidity current could result in changing current directions and that the ripple directions would not correspond with the sole markings. There need not of course be exact correspondence as the erosive phase that gives rise to the bottom structures may differ in direction from

from the following depositional phase. This can be due to change in the direction of the current relative to the slope of the bottom as it decelerates, or to general fanning out of the current during flow. In the few instances where current directions could be determined from ripples on the top of a turbidite they indicate a direction of transport roughly parallel to that shown by the bottom structures, thus it is possible they were produced by one of the above mechanisms.

The author considers it most likely that the ripples were produced by a relatively sediment free current, this giving ripples that were not quickly stabilised in contrast with the c division ripples which were apparently quickly stabilised. The experiments of Simons et al. (1963) already cited indicate that it may be possible to distinguish ripples formed from a muddy current and those from a clear water current on the bases of shape and prevalence of ripple lamination truncation.

It is also possible that the ripples were formed by a current unconnected with the turbidity current. An ordinary bottom current within the basin could easily cause erosion of the d and e divisions of a turbidite and rework the sandy part into ripples. In this case the currents need have no relation in direction to any

turbidity currents. If this mechanism has been operative in the present area the currents must have been intermittent in nature.

In two cases (Unit C, Thorncliff 422 and Croker 354) trails left by surface crawling animals were found crossing ripples on the tops of rippled beds. This would seem to indicate that after their formation the ripples were left exposed on the sea floor for a little while before being covered by mud. The preservation of the trails at least indicates the intermittent nature of the currents causing the rippling, but it is impossible to tell whether the reworking was due to a non-depositional phase in a turbidity current, a clear water tail to the turbidity current, or normal bottom currents of an intermittent nature.

Walker (1965 p.18) has observed cases of isolated ripples in mud from the Westward Ho! Formation (U. Carb. N. Devon. De Raaf et al. 1965) which he considers to have been formed by the reworking of a turbidite by the tail of the turbidity current. Similar structures observed by Hsu (1964) have been attributed by him to reworking by ordinary currents.

Simons et al. (1963) show that ripples become stabilised when there is a proportion of fine grained material in the flow. The erosive bases to the ripples

and the truncated laminae observed in them show that they were not a static bed form but were moving. Thus it could be argued that they were formed under relatively clear water conditions. There is still no evidence as to whether the clear water current causing the rippling was initiated by the turbidity current and followed it rather like a wake, or whether it was an ordinary bottom current unconnected with the turbidity current. It is also possible that a large turbidity current could induce a normal bottom current that carried on moving long after the turbidity current had deposited its load.

ORIGIN OF THIN BEDDED PROXIMAL DEVELOPMENTS.

The thin bedded proximal developments already referred to are characterised by strong discontinuous tool marks and some flute moulds. The beds are generally thin less than 10cm. and may be represented by isolated ripples in mudstone, see Fig. 2.1Eb. Erosion of previously deposited sand by succeeding currents is frequent, and the beds are set in green shale mudstone. This situation contrasts markedly with the typical proximal beds already described as deposits of immature turbidity currents.

The sole structures indicate extreme turbulence of the eroding current which must have been strong to

produce the large prod marks seen. It is possible that the turbidity current transported nearly all its material over the area of the thin bedded proximal developments and did not begin deposition until later, perhaps when the slope flattened out. This implies that the current might actually have been gaining considerable material by erosion of the substrate as it passed.

Another possibility is that the turbidites were deposited normally and were eroded by ordinary bottom currents. These currents could rework thin turbidites into isolated ripples and could cause the rippling of the tops of the deposited sand beds as described in the previous section.

There is no evidence in the sections studied to show the relationship between the proximal deposits and these thin bedded sequences. Possibly the thin bedded sequences occupied relative 'highs' on the sea floor which were not affected by the immature currents depositing the proximal beds.

2.1I

CONCLUSION ON THE ORIGIN OF THE PROTOQUARTZITES.

The protoquartzites show all the features of typical turbidite formations and are considered to have been deposited mainly by turbidity currents. General 'Proximal' and 'Distal' sequences can be recognised and are considered to reflect increasing maturity in horizontal and vertical sorting in the turbidity current as it flowed out into the basin.

There is evidence of post depositional modification of the tops of the turbidites which causes elimination of the d and e divisions and reworking of the top of the sandy part of the bed into ripples. The reworking is possibly due to an ordinary bottom current within the basin or to a clear water 'tail' current following the turbidity current.

It should be stressed that it is the complete assemblage of features present in the protoquartzites which indicates that they are turbidites. This assemblage consists of bottom structures of tool marks, scour marks and load structures; the consistent internal sequence of structures a to e with base out types; the grading of the beds above a sharp erosional base; extensive continuity of beds and evidence that they can be traced from a proximal to a distal region and the presence of a derived flora (land plants in this

case). The features most conspicuously absent are large scale cross stratification, lensing of beds and a shelly benthonic fauna.

The term flysch has been applied to many turbidite formations which usually occur on a much larger scale than the present occurrence. The 16 characteristic features of flysch given by Dzulynski and Smith (1964 p. 248-9.) are all obeyed by the protoquartzites apart from No.13 which states that ripples are scarce on the tops of sand beds. This is not true in the present area.

Most flysch filled basins are on a very large scale and the present occurrence is on a relatively small scale and is not technically comparable with the great flysch filled troughs found associated with mountain chains. The situation in north Staffordshire corresponds with the autogeosyncline of Kay (1951) as described by Holdsworth (1963).

It is best to regard the north Staffordshire area as a basin in E1 - E2 times that was partly filled with the turbidites, and not to apply the term flysch to every turbidite bearing sequence.

2.1 J

DIRECTIONAL AND REGIONAL FEATURES OF THE PROTOQUARTZITIC
TURBITES.Current Direction Data.

In the three units of protoquartzites studied, current directions based on sole structures were measured wherever possible. Readings from inclined bedding surfaces were corrected by rotation of the bed to the horizontal about its strike.

The directions were recorded separately for each of the units A, B, and C but insufficient readings were obtained (due to poor exposure) to warrant separate treatment of the results. All three units showed transport from the south and the range of variation obtained in the units was similar except in the Thorncliff area. On the paleocurrent map for the protoquartzites (Fig. 2.1Ja^{*}) all the readings taken at one locality have been averaged to give the prevalent current direction at that point. The number of readings being indicated as well as the unit (A, B or C) on which the measurements were made, when this could be determined with certainty.

Groove and bounce marks frequently showed the line of the current but not the direction, this being determined in most cases from the prod moulds since flutes are scarce in the area. In all cases prod moulds indicate current transport from the south.

* Figure in folder.

The only exception to this being in the top part of Unit C at Thorncliff and Easing where a few beds indicated transport from a westerly direction.

Fig. 2.1Ja is based on over 200 current direction readings. Where more than one reading had been taken from an individual bed at a locality these were averaged and the value obtained used along with readings from other beds at the locality to find the average current direction for the locality. Thus no one bed was given extra significance because it bore clear sole markings.

In the western part of the area the average current directions indicate currents flowing about due north. In the eastern part directions are more variable but the bulk of the readings indicate currents flowing from about 20° west of south. In the extreme east of the area currents have been noted coming from just east of south. It is not certain whether these variations are really significant or due to the small number of observations involved. In the Thorncliff - Easing area there is a marked difference in trend between the directions obtained from units B and C, due to the presence of a few westerly derived beds in Unit C. Unit B shows current directions from almost due south whilst it is those of the A and C units which come from west of south.

The regional significance of the current direction distribution will be considered when the areal features of the proximal and distal beds have been discussed.

Areal distribution of Proximal and Distal type beds.

Groups of Localities showing similar developments.

- | | | |
|--|---|--------------------------------------|
| 1. Endon, Shirkley and Sprink. | | Proximal. |
| 2. Gun Hill. |) | |
| 3. Croker Hill, Thorncliff. Easing. |) | |
| 4. Hurdlow. |) | Mixed Proximal and distal sequences. |
| 5. Pyecloough, Oakenclough. Sunnydale. |) | |
| 6. Blake, Elkstones. Martinslow. |) | Distal. |
| 6. Stannery, U. Dove. Waterhouses. | | Shale-mudstone. |

PROXIMAL.

1. Endon, Shirkley and Sprink.

All these localities show strong developments of the typical proximal type already described and are interpreted as being the nearest available exposures to the source area of the protoquartzites. The proto-quartzites are thick and lack good internal grading and lamination, beds 50 cm. thick are of frequent occurrence, and pebbles are common. Occasional thin bedded developments are present and these have been described in section 2.1E.

MIXED PROXIMAL AND DISTAL.

2. Gun Hill.

The beds exposed in Gun Stone Pits Loc. 402 are allied to the proximal developments but are generally better sorted and show more grading and lamination than the proximal developments of group 1. Exposures other than those in the Stone Pits are very poor and as the pits have obviously been excavated in a part of Unit C containing the thickest sandstones, a false impression is probably attained of the average strength of sandstone development at this locality. It does however appear that some very thick beds (greater than 1m.) occur in this area and current directions indicate they are derived from the south. These beds were deposited from slightly more mature currents than those which affected group 1 localities. The thickness of the beds is surprisingly large and this area must have been a favoured site for deposition from the turbidity currents possibly due to an abrupt change of slope of the basin floor. The majority of beds are proximal in aspect but do not have all the typical proximal features and the increase in importance of lamination and better sorting indicates a more distal deposit than that of group 1 localities.

3. Croker Hill Area.

Numerous exposures around Croker Hill show the wide variation to be found within the protoquartzites in any one area.

At Loc. 351 the base of Unit C is seen to consist of thin bedded protoquartzites usually less than 10 cm. thick and often displaying rippled tops. Following this relatively weak entry of the proto-quartzites, is a strong development of beds up to a 130 cm. thick and often with welded contacts as seen in the quarry at Loc. 355. These beds resemble superficially the proximal beds of area 1 but inspection shows them to be far better sorted and quartz grains rarely exceed 2 mm. in diameter. Quartz pebbles are absent and clay pebbles very rare. The internal structures cannot be observed satisfactorily due to the state of exposure and the staining of all joint surfaces in the beds. The beds are often loaded and strong groove marks are present on one bed.

Intermediate in strength between these developments are those seen at Locs. 359 and 360. At these places protoquartzite developments are intermediate between the two extremes described above. A few beds only exceed 50 cm. in thickness, the average being 10-20 cm. (e.g. 14 cm. average thickness for sand beds in Unit B

at Loc. 359). Bottom structures are well developed and consist of grooves, which are often long and continuous, bounce moulds and prod moulds. Flute moulds occur rarely on the bases of some of the thicker beds. The internal structures appear to be well developed with the divisions a to e present in suitable beds, and the normal base cut-out sequences present in others. Beds with rippled tops are of frequent occurrence. At Loc. 359, of 25 bed tops exposed 17 are rippled and only 8 graditional to shale-mudstone.

At Fox Bank Quarry (360) a development presumed to be in Unit B shown that the turbidite beds are themselves grouped into units separated by shale-mudstone containing only a few thin turbidites, this feature is not often seen in the small exposures usually available. The turbidites at this locality are visibly grouped into 5 units containing from 11 to 24 sand beds in a unit. The lower beds in the unit are usually the strongest. This feature will be referred to later.

The general impression gained from the Croker Hill area is that of one over which turbidity currents of a mature nature were flowing and depositing their load. The beds deposited being variable in thickness according to strength of the currents reaching the area. Rippled tops to beds are frequent indicating action of

clear water tails to the turbidity currents, or normal bottom currents (see section 2.1H).

The beds exposed in the Hawkshead quarries Locs. 370-371 are different from those seen in the rest of the Croker Hill area and more closely resemble those seen at Gun Hill. The exact stratigraphic age of these beds cannot be determined but they indicate strong currents from the south depositing groups of thick protoquartzites with internal structures lacking or poorly developed. Slumping also affected this area as shown by the beds at Loc. 370 which indicate slumping from the west.

The three developments described so far are those from the western part of the area where current directions (Fig.2.1Ja) consistently indicate transport to the north. The protoquartzite beds show a change in this direction from proximal type to a predominantly distal type. No exposures are known north of the Croker Hill area and so it cannot be ascertained with certainty where the beds finally die out. Strong mature turbidity currents were certainly present flowing northwards in the Croker Hill area which, given a favourable slope, could have continued for several miles at least.

3. Easing and Thorncliff.

Turning now to the eastern part of the area the first noticeable point is that^{at}_A the latitude of Shirkley and

Sprink in the western area the developments in the east are markedly different. They are more distal in nature than those in the west.

The beds exposed at Easing and Thorncliff indicate a development similar to that already described from the Croker Hill area. Individual beds are usually less than 50cm. thick and with the exception of a single bed at Thorncliff quartz pebbles are absent. There is a strong development of protoquartzite beds at the top of Unit C at Thorncliff 422 that has been quarried in the past. This is similar to the development seen at Loc. 355 Croker. Below this development thinner beds are present often with extensively rippled tops. These ripples are well exposed on an extensive dip slope forming the stream at Loc. 422, and some trails up to 5cm. wide are present on the rippled top of the bed. It is probably purely coincidental but the only other place where such trails have been observed is at Croker Hill (354) also near the top of Unit C of the protoquartzites.

Exposures at Easing are poor but there is evidence of a relatively strong protoquartzite development at the top of Unit C and another near the base, each with protoquartzites of maximum thickness c.50cm.

Isolated sand ripples in shale-mudstone are seen occasionally at these localities and are similar to those in the thin bedded proximal developments.

4. Hurdlow.

In the Hurdlow area very few protoquartzite beds exceed 50cm. in thickness and the vast majority are less than 20cm. thick. The grouping together of the protoquartzite beds is again seen here as it was at Fox Bank Quarry Loc. 360. Within Unit C there appear to be three groups of thicker protoquartzite beds present, whilst between these three groups shale-mudstone is the predominant lithology. At Thorncliff the strong protoquartzite development at the top of Unit C had been quarried and contained more and thicker beds than those seen at Hurdlow. The equivalent part of the succession is exposed at Locs. 010, 015, and while there are several protoquartzites up to 50cm. thick they are not so numerous as at Thorncliff Loc. 422. In this part of the succession there is a definite reduction in the number and thickness of the protoquartzite beds from Thorncliff to the Hurdlow area.

Another strong protoquartzite development occurs near the base of Unit C and is exposed at Locs. 009 and 022 where the base of the development is about three metres above the E2a2 faunal band. The other strong development occurs roughly in the middle of Unit C and is exposed in the stream bed between 009 and 010. These stronger developments contain only a few beds resembling the proximal developments already described and the Hurdlow area represents the last appearance of these

internally structureless beds.

Beds with rippled tops are frequent in the Hurdlow section, particularly just above the E2a2 faunal band in the base of Unit C (Loc. 022) and also at the top of Unit B where a rippled surface forms the stream bed at Loc. 001. The top part of Unit B contains a strong development of protoquartzites seen near Loc. 008a. Unit A is exposed at Loc. 019 and Loc. 012 where evenly bedded protoquartzitic turbidites are exposed which generally do not exceed 30cm. in thickness and are mainly base cut out beds. Groove, prod and bounce moulds are frequent and so are unroofed burrows. A larger proportion of these beds have gradational tops than those in Unit C. One exceptional bed 40cm. thick at Loc. 019 is due to welding of the deposits of at least two currents.

The general aspect of the beds exposed at Hurdlow is close to that seen at Thorncliff but there is demonstrable thinning to the north. The majority of the beds appear to have been deposited by reasonably mature turbidity currents since they have developed in them the typical internal structures. There was a great range in the strength of the currents supplying the area, but it is again noticeable that the thicker beds are grouped together rather than scattered randomly through a turbidite unit. From the Gun Hill type of development

to the type seen at Hurdlow there is a progressive reduction in proximal type beds which are internally structureless and an increase in distal beds with well developed internal structures. This is in agreement with the findings of Walker (1967 p.35) who notes a marked increase of ae type beds in proximal areas.

DISTAL.

5. Pyecloough; Oakenclough; Sunnydale; Blake; Elkstones; Martinslow.

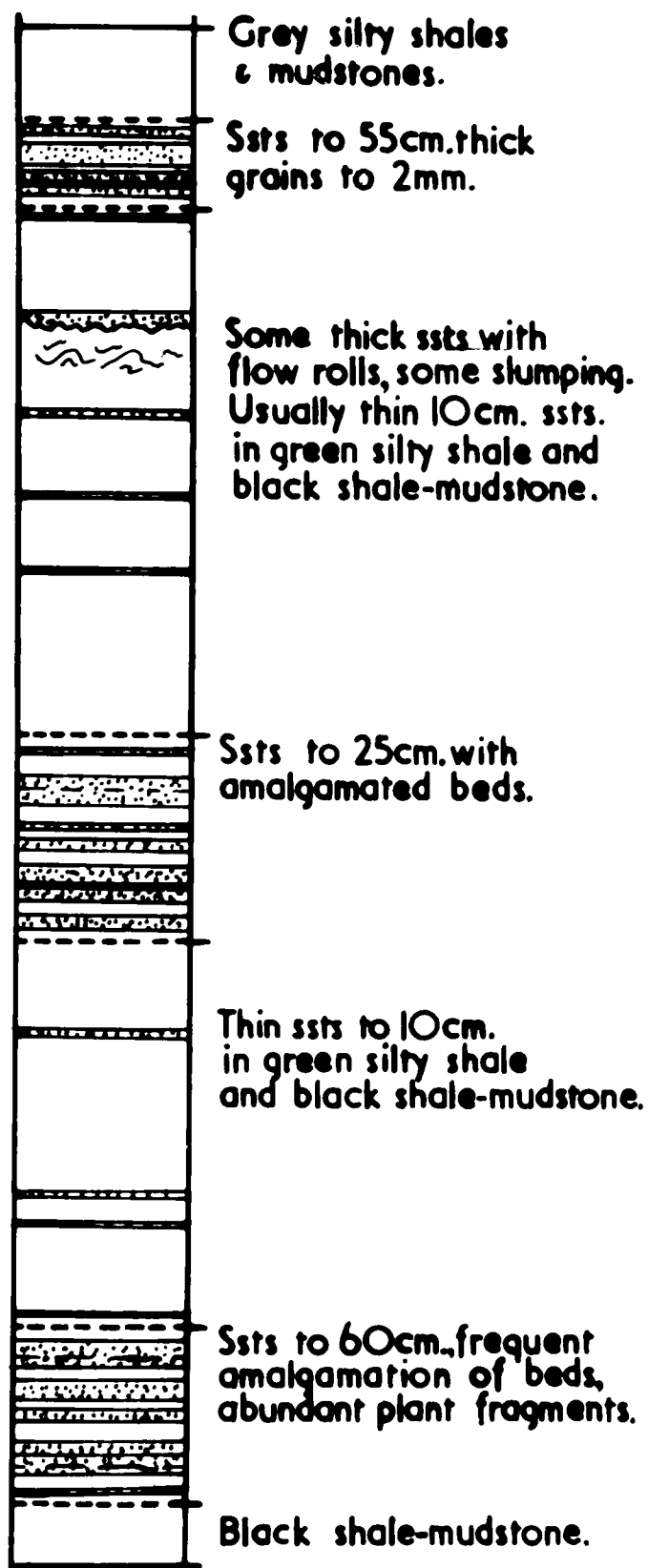
All these localities display distal type sequences of protoquartzites. There is still room for considerable variation in the aspect of the beds, but in general beds greater than 20cm. thick are absent or rare. The vast majority of the beds present are base cut-outs of bcde, cde or de types with some b-de type beds as previously described.

As an example of this most distal expression of the protoquartzites the C unit at Pyecloough will serve as an example. Shale-mudstone is the dominant lithology. All protoquartzite beds are less than 20cm. thick and most are less than 10cm. They occur grouped together in two places within the C unit thus presenting a similar appearance to that observed in the more proximal development at Hurdlow (Fig.2.1Jb). Small groove and prod moulds are the most frequent bottom structures and

Fig 2.1Jb

COMPARISON OF UNIT C PROTOQUARTZITES

HURDLOW



PYECLOUGH

Black shale mudstone with siderite nodules.

Ssts to 4cm. thick with siderite beds.

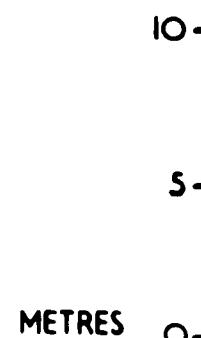
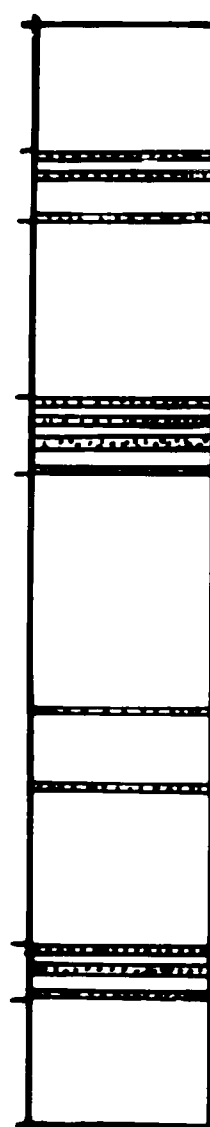
Shale mudstone with green silty beds.

Ssts to 10cm. thick few with rippled tops.

A few thin ssts with many green silty beds in shale mudstone with siderite beds & nodules.

Ssts to 8cm. thick, nine present in one metre.

Shale mudstone with siderite nodules.



unroofed worm burrows are also frequent. No quartz pebbles are present but a few shale pellets occur in the thicker beds. Plant fragments are abundant on some lamination planes.

These developments are the deposits of weak turbidity currents which had slowed down and deposited most of their load before reaching the area. They represent the most distal developments seen of the turbidity currents which deposited the protoquartzites. In the distal areas the proportion of beds with gradational tops is greater than in other areas. In logs of protoquartzite developments at Pyecrough (Units A and C) and Oakenclough (Unit A), only 14 beds out of 81 lacked gradational tops and were top truncated. The average thickness of the protoquartzite beds is less than 10cm. but this is a rather meaningless statistic since the beds are not randomly distributed through the section and the sections measured tend to be in the stronger developments where exposure is better.

DEPOSITS BEYOND THE RANGE OF THE TURBIDITY CURRENTS.

6. Upper Dove Valley - Stannery; Waterhouses.

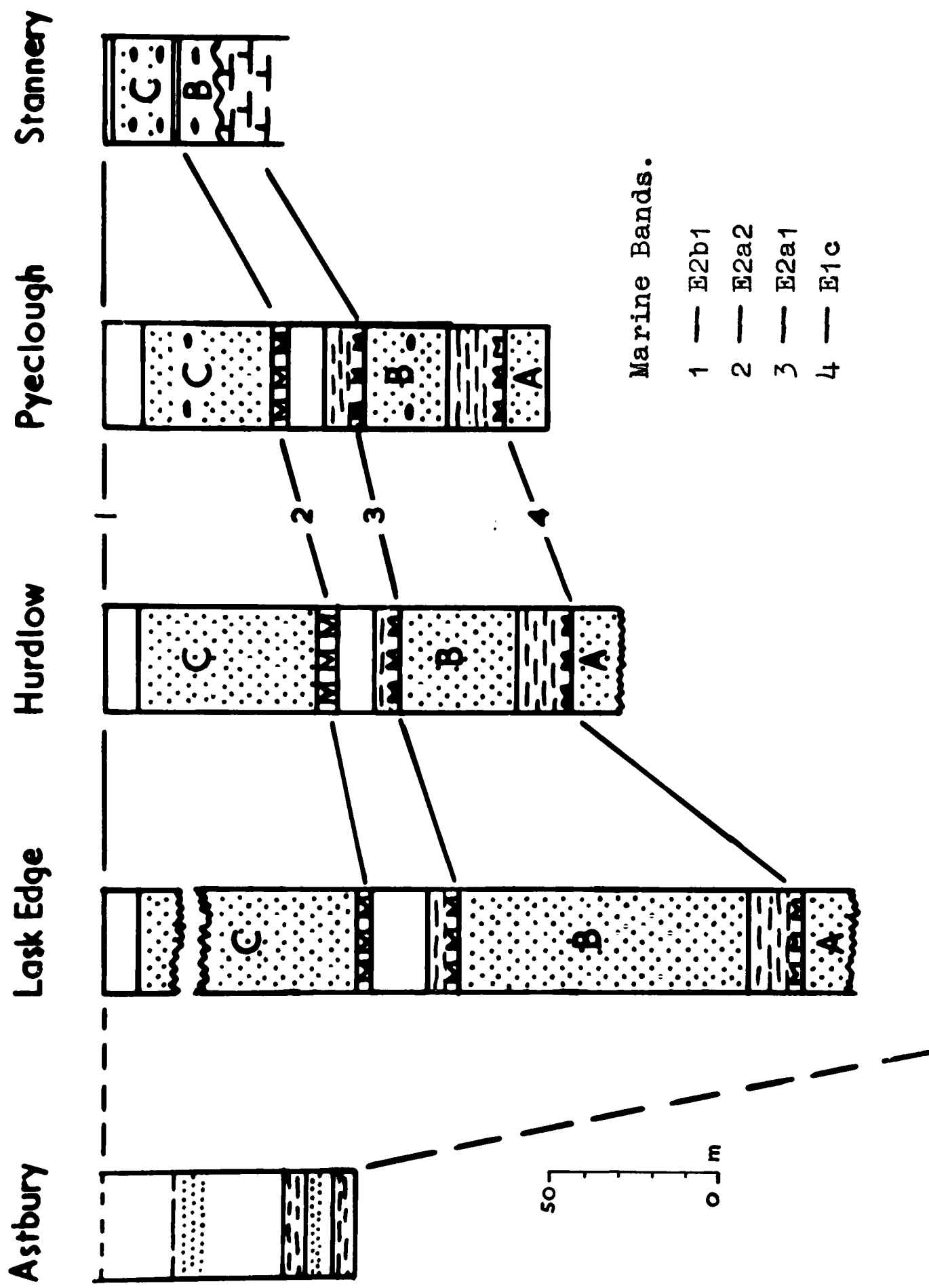
The exposures in the Upper Dove are situated very close to the point of unconformity of the Arn^sbergian onto the Viséan reef limestones. The actual lateral distance of the exposures of the beds representing Unit C from the surface of unconformity varies from about

50 to 200 metres. Unit C is the only one seen at this locality and it is represented almost exclusively by shale-mudstone with siderite nodules. Only one 3cm. bed of sand has been seen in the section (which is reasonably exposed) and a few green silty beds are present similar to type e division developments. It is presumed that the bulk of the turbidity currents did not reach this area at all, having died out between Pyecloough and Stannery — a distance of under two miles.

At Waterhouses Loc. 561 the E2a2 faunal band is exposed and above it there are 15 metres of shale-mudstone with some siderite beds and nodules but no beds of sand or silt are present. This is therefore a similar development to that seen in the Upper Dove. None of the exposures 562-565 show any protoquartzites, neither do the exposures in the railway cutting near Loc. 622 where the E2a2 faunal band is again exposed. It is clear that this area is identical in aspect to the Upper Dove and the proximity of the Carboniferous limestone outcrop is also significant. Here the junction of the Visean and Namurian is further from the exposures and is possibly conformable since a good sequence of basal E1 goniatites has been collected in Waterhouses Railway Cutting (Morris 1967) and a form of E. pseudo-bilingue (E1b) is present in the Waterhouses School section Loc. 601. The relation of E2 to these beds and

Fig 2.1Jd

THICKNESS VARIATIONS OF PROTOQUARTZITE UNITS A, B AND C.



to the limestone cannot be demonstrated. It seems probable that the whole of the succession in E zone is generally lacking in coarse grained lithofacies except for a small development of thin muddy sandstones in E1 exposed in the section at Loc. 601.

The distribution of areas of protoquartzite deposition with approximately the same aspect, or degree of proximity is shown on Fig.2.1Jc*.

THICKNESS VARIATIONS IN PROTOQUARTZITE UNITS.

It has already been shown that the individual protoquartzite beds show a general reduction in thickness away from the source area. This feature being most marked when the beds are traced from Thorncliff to the Upper Dove. The A B and C units of protoquartzites show a similar reduction in thickness in the same direction. This is largely due to the reduction in number and thickness of the turbidite beds but also due to a reduction in the grain size of the normal bottom sediment. The variation of the 'normal' sediment will be discussed later (section 2.2). Fig.2.1Jd is a comparison of various sections to illustrate the general features of the thickness variations.

Localities where the thicknesses of the units can be accurately measured are unfortunately few in number since the required sections have to expose the top and

* Figure in folder.

bottom of the unit and sufficient of the unit to show that minor faults and folds are not present.

The A unit is 12 metres thick at Pyecclough and 12.5 metres at Blake, these two localities showing very similar developments. In the Oakenclough Brook its thickness is 17 metres and in the Hurdlow area it can be demonstrated to be at least 15 metres thick with the base not seen. It is probably thicker than 17 metres as the turbidites are thicker than in the Oakenclough section. At Thorncliff the thickness appears to have increased again with thicker turbidites present but accurate measurements cannot be made here.

The thinnest development of Unit B measured is 14 metres at Elkstones close to the eastern edge of the basin. Here the protoquartzites are very poorly developed. At Pyecclough the thickness is 21 metres and at Blake 22 metres. These localities again showing similar developments and almost identical thicknesses. At Hurdlow the B unit is 30m. thick with a strong development of protoquartzites at its top. The exposure in Fox Bank Quarry 360 (Crocker Area), presumed to be of Unit B, shows at least 28m. of protoquartzites and as neither the top or bottom of the unit is exposed it is probably thicker than that at Hurdlow. In the proximal area at Sprink the B unit is at least 80 metres thick and could be much thicker — possibly as much as 120 metres. Thus there is

considerable thinning away from the proximal area in Unit B.

The thickness of Unit C at both Pyecrough and Blake was measured as 37 metres thus continuing the striking similarity shown by these localities in the A and B units. Further from the source area the thickness at Sunnydale is at least 21 metres but probably not much more than this figure. In the Upper Dove where the protoquartzites have died out only 12 metres of shale-mudstone with siderite nodules represents Unit C.

Nearer the source area than Pyecrough and Blake a thickening of the unit can be demonstrated. At Oakenclough the C unit is about 43 metres thick and at Hurdlow it measures 50 metres. Further south in the Thorncliff-Easing area the total thickness could not be measured but the individual groups of turbidites in the C unit have increased in thickness and so the unit as a whole is probably thicker.

A full exposure of Unit C is lacking in the Croker and Gun Hill areas, and in the proximal area at Upper Shirkley 36 metres of the base of the unit can be seen and from ground features it appears to be over 70 metres thick.

Although it would improve the general picture to have more accurate thickness measurements for the A, B and C units, those that have been obtained are reasonably accurate and show an interesting pattern. There

appears to be a strong correlation between the thickness of the units and the general proximal-distal aspect of the turbidites. This is well shown by the Pyecrough and Blake sections which are of almost identical thickness and show similar developments of turbidites. A general increase in thickness from these localities through Oakencrough, Hurdlow and the Thorncliff areas is also apparent. The thickness of the units in the latter area being approximately equivalent to those in the Croker Hill area.

REGIONAL INTERPRETATION OF THICKNESS AND PROXIMALITY OF TURBIDITES.

When this information of thickness variation is considered alongside the evidence presented in the previous section on the variability of the protoquartzites over the area, and the current directions, a diagram can be drawn showing areas of roughly equal development of the protoquartzitic turbidites (Fig.2.1Je^{*}). Crude though this diagram is, it is readily apparent that the areas of equal development form bands which are markedly oblique to the current directions, especially in the east of the area.

If the floor over which the currents were flowing was of even slope in all directions away from the source then areas of similar turbidite developments would be expected to form bands perpendicular to the current

directions. The bands would maintain a roughly constant distance from the source of the currents and parallel to the bottom contours of the area (Fig.2.1Jf).

If the slope away from the source area was a gently dipping plane then a pattern as seen in Fig.2.1Jg would possibly be expected with bands of similar turbidites forming crescent shaped areas with the noses of the crescents pointing downslope.

The pattern which arises from the observations in the present area does not conform to either of these simple patterns although there is some similarity to that shown in Fig.2.1Jg.

The distribution of the bands of equal strength turbidites (Fig.2.1Je) seems to indicate that the main axis of flow was from south to north in the western part of the area for it is here that the change is most gradual in a downcurrent direction giving bands of similar aspect turbidites. In a NNE direction in the east of the area the bands are oblique to the current directions and closer spaced reflecting the manner in which the turbidites die out as they approach the old limestone massif probably due to a slope back into the basin near the massif margin.

The lack of strong turbidites in the Waterhouses area reflects a relative high over which the currents did not flow. This is also borne out by the weak devel-

Fig 2.1Jf

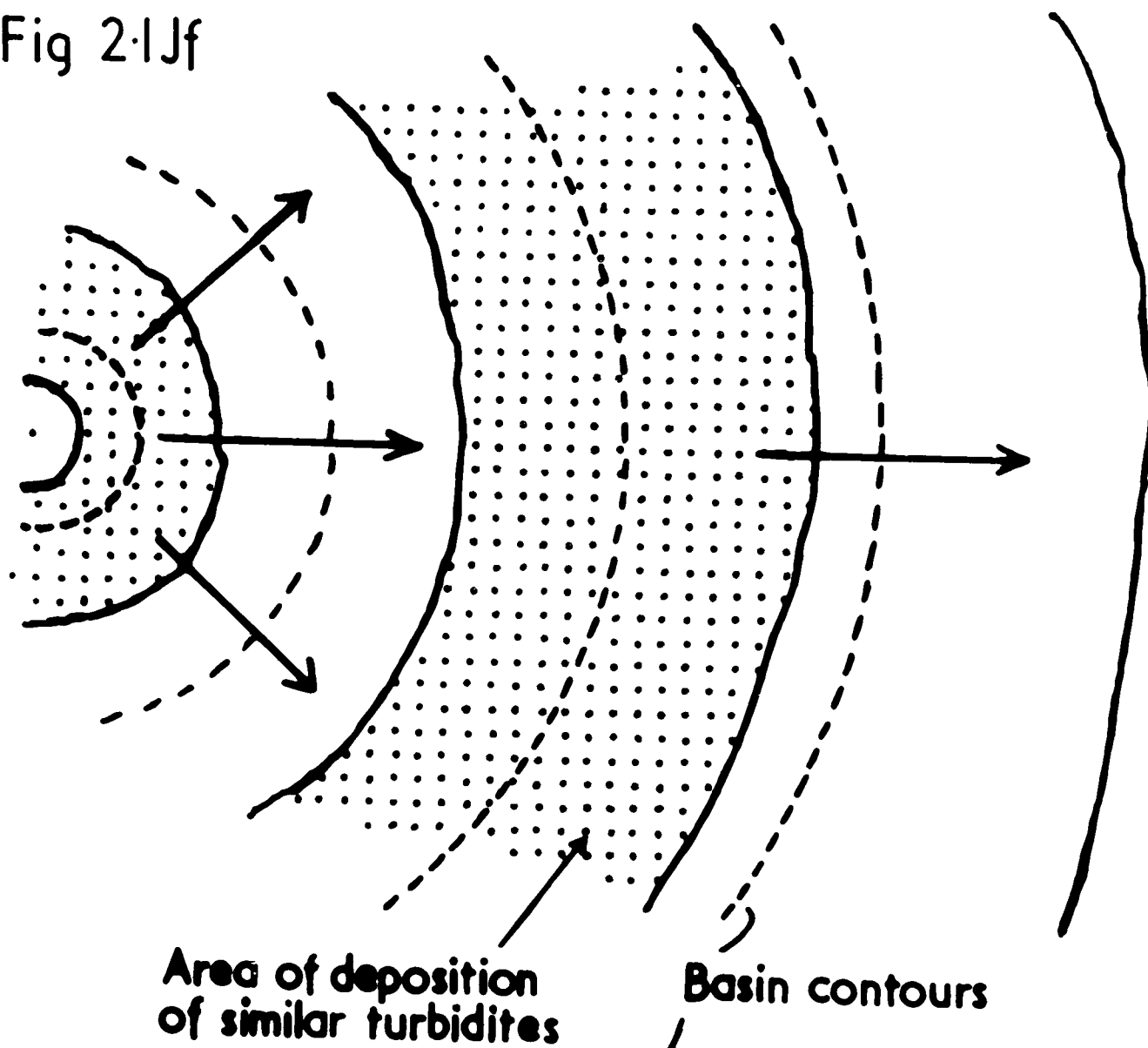
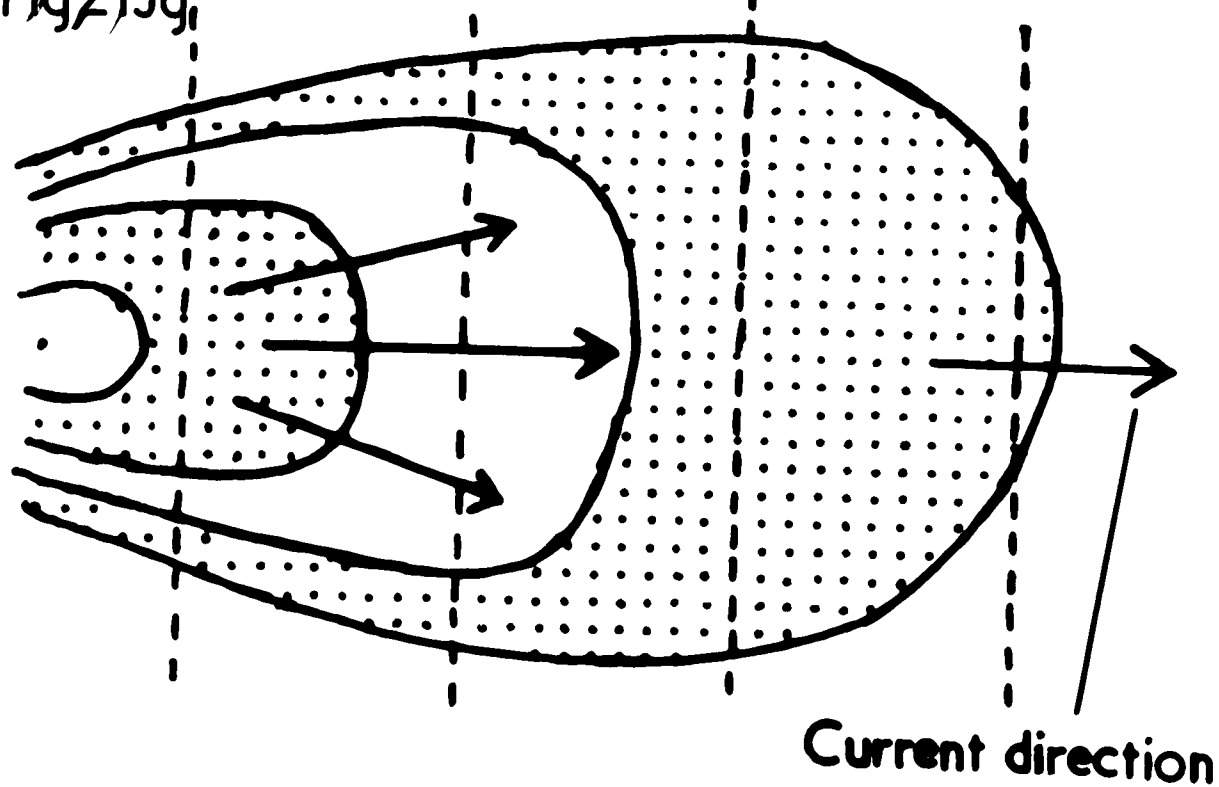


Fig 2.1Jg



opments at Winkhill-Martinslow. The bulk of the currents seem to have flowed northwards from south of the Endon-Shirkley-Sprink area, but a proportion fanned out to give the weaker developments seen in the east of the area.

Exposures to the west of the main axis of flow at Astbury (section 3.1) show that the turbidites were absent in this area and this feature combined with the slump features at Croker Hill seems to indicate that the basin had a steep western edge. This steep western slope could serve as a source for the slump deposits of Loc. 370 as well as stopping the turbidites reaching Astbury. The full significance of the Astbury section is considered in section 3.1.

2.1. K

RELATIONS BETWEEN MEASURABLE PARAMETERS IN THE
PROTOQUARTZITE UNITS.

19 sections of the protoquartzites, 9 distal and 10 proximal were logged during the study, the number of turbidites in each log ranging from 10-45. The shortness of the exposed sections makes detailed mathematical treatment of the results somewhat meaningless but certain features do emerge and are worthy of note.

Because of the difficulty of recognising the internal divisions of the turbidites in the field due to silica cementation the proportions of different types of beds involving the a, b and c divisions are not known, and would be meaningless if expressed for the area as a whole.

From the measured logs (Fig.2.1Ka) the proportions of protoquartzitic sand, representing the abc divisions, green silt (de divisions) and basinal shale-mudstone were calculated and also the average thicknesses of the beds of each type. Fig.2.1Kb shows the three components plotted as a triangular diagram. The proximal developments poor in shale-mudstone plot along the Sand-Green Silt join and the distal points are scattered in the centre of the field. No points indicating sequences with greater than 75% basinal shale-mudstone have been plotted although these exist, (e.g. in the Upper Dove,) and

Fig.2.1Ka

Locality and Log Number	% sand abc divisions	% Green Silt de divisions	% Shale- mudstone	Ave. th. Sand (cms)	Ave. th. Green Silt (cms)	Ave. th. Shale- Mudstone (cms)	Turbidites top truncated %	Sequence Type
1. Pyeclough A2	7.8	49.6	42.6	0.64	1.94	1.66	0	D
2. Pyeclough A1	18.1	47.5	34.3	2.9	2.2	1.6	0	D
3. Pyeclough C1	41.8	28.7	29.5	5.2	2.5	2.6	10	D
4. Pyeclough C2	54.2	43.7	2.1	7.1	6.3	3.0	90	D
5. Oakenclough A	60.9	29.7	9.4	6.2	3.0	1.4	21	D
7. Croker B	81.9	15.7	2.4	14.1	3.3	3.0	68	D
18. Hurdlow A	32.3	42.2	25.5	4.9	4.6	3.66	22	D
8. Hurdlow A1	27.3	43.9	28.8	3.94	4.00	3.07	15	D
9. Hurdlow A2	37.4	40.1	22.5	5.93	5.34	4.33	32	D
19. Shirkley C	95.7	4.3		31.2	4.3		100	PT
6. Shirkley C	65.5	34.5		9.9	7.5		96	Pt
10. Endon Mill 4	72.6	27.4	-	8.18	3.19	-	100	PT
11. Endon Mill 3	31.6	61.6	6.8	4.78	4.44	0.87	41	Pt
12. Endon Mill 2	50.5	47.4	2.1	4.57	3.99	0.83	78	Pt
13. Endon Mill 1	58.7	35.8	5.5	4.26	2.72	1.2	67	Pt
14. Sprink 1	36.0	27.9	36.1	2.85	1.78	3.74	41	Pt
15. Sprink 2	25.2	34.6	40.2	1.95	2.09	6.54	51	Pt
16. Sprink 3	81.4	17.4	1.2	22.5	6.32	1.70	79	PT
17. Sprink 4	80.7	19.3		12.7	3.82		100	PT

D = Distal PT = Thick Bedded Proximal

Pt = Thin Bedded Proximal

consist of basinal shale-mudstone with occasional green silty beds and sand practically absent. Sequences composed of green silty beds of d-e type would probably not be recognisable as such until there was sufficient black shale-mudstone or sand present to delimit individual beds, thus it is not likely that sequences composed of more than 70% green silt (d-e divisions) would be recognised and no measured sections fall in this range and only one, a proximal development, has more than 50% green silt. The expected field of variation is indicated in Fig.2.1Kb.

When the percentages of the three components are plotted against their average bed thicknesses some interesting features emerge. There is a strong correlation between % sand and the average thickness of the sand beds Fig.2.1Kc (linear correlation coefficient $r=.83$) this holds generally for both proximal and distal sequences although the distal sequences show a better relation due to the erosion and amalgamation present in the proximal sequences. The graph shows that as the percentage of sand increases there is a strong tendency for beds to become thicker rather than more numerous. The relation appears to be roughly linear up to 70% sand content but above this value there is a more rapid increase in sand bed thickness (see logs 16 and 19 not on graph).

A plot of green silt average bed thickness against % green silt (Fig.2.1Ke) shows a general distribution field rather than a definite trend. The average bed thickness remaining roughly constant between 1.7cm and 7.5cm. and the majority lying between 1.7 and 4.5cm. The linear correlation coefficient $r=.025$ indicates that the bed thickness is independent of % green silt. Similarly a plot of basinal shale-mudstone average bed thickness against % basinal shale- mudstone (Fig.2.1Kd) shows no clear trend ($r=.495$). There is a concentration of points on Fig.2.1Kd representing sequences with less than 10% dark shale-mudstone with average bed thickness less than 4cm. There is a slight increase in average shale-mudstone thickness associated with distal sequences. This is to be expected since erosion of shale-mudstone would have been more prevalent in the proximal areas.

In the field there appear to be far more top truncated beds in the proximal areas than in the more distal regions. This is largely a reflection of the proportions of sand and shale-mudstone present and a plot of % sand against % beds top truncated (Fig.2.1Kf) shows that there appears to be a general relation between the two parameters and confirms that sequences with a low percentage of sand tend to contain fewer top truncated beds.

There is however a poor linear correlation between these two sets of data ($r=.45$) and this can be explained

as follows. The data represent both proximal and distal sequences and when individual points are identified it is found that proximal sequences contain a greater proportion of top truncated beds than distal sequences when the sand % is roughly the same. The only exception is log 4 which is a short log (including only 10 turbidites) in an atypical part of Unit C at Pyeclough. Thus in the north Staffordshire basin area it appears that as a general rule top truncation is more prevalent in the proximal areas than the distal. As would be expected there is a good linear correlation ($r=.82$) between the % of shale-mudstone and % top truncated beds (Fig.2.1Kg) the only points plotting far from the main trend being two thin bedded proximal sequences with a high proportion of beds with rippled top surfaces.

The ABC proximality index of Walker (1967) as used by Walker and Sutton (1967) on the Upper Devonian of New York cannot be applied to the present area since in the field it is not always possible to be sure of the internal structures present in the turbidites, particularly in the case of the c division. Walker does not include consideration of the d and e divisions which have been recognised during the present study. The ABC index, being based on the frequency of occurrence of beds beginning with divisions a, b or c does not differentiate between ae beds of typical proximal aspect and abcde beds deposited by mature currents often in a distinctly more

distal environment. Walker (1967 p.35) plots ABC index against % ae beds and shows that these are far more frequent in the more proximal areas as was observed in the present area.

Walker (1967 p.33) shows the relation between the ABC index and the average thickness of the sand beds and finds only a gradual increase in average sand thickness in the range of ABC index $P_1 = 10\%-60\%$ but a rapid increase above $P_1 = 70\%$. In the present area the lack of complete measurable sections through units A, B and C makes meaningful calculation of ABC indices impossible since it has been demonstrated that the strong turbidites tend to be grouped together within a unit and calculation based on available sections would give non-typical values for a particular area. It would be useful to compare some results plotted against the ABC index with the same parameter plotted against sand % for a suitable section. This cannot be done in the present area without collection of all sand beds for cutting to reveal the structures.

RELATION OF GREEN SILT TO SAND THICKNESS FOR INDIVIDUAL TURBIDITES.

Fig.2.1Kh shows the relative proportions of green silt (de divisions) and sand (abc divisions) for about 100 distal beds which are not top-truncated. It reveals that the thickness of the d+e divisions rarely exceeds

6cm. and is independent of sand bed thickness. This variation in the thickness of the d and e divisions is quite small and indicates that the finer material was deposited fairly evenly on the top of the lower divisions in the distal environment, and continued to be deposited beyond the range of the a to c divisions of the turbidite. It thus seems likely that in the case of top truncated beds where the truncation is due to later rippling of the bed, the current that caused the rippling would have had to erode up to 6cm. of green silt and a few sandy laminations in the case of an ordinary current flowing after deposition from a turbidity current had ceased, before a clean reworked rippled sand top to the turbidite could be obtained. Alternatively, if the rippling was due to the turbidity current tail the material would have been kept in suspension and transported to more distal environments before reworking took place.

While these simple treatments of the data obtained from the turbidites do little more than to express diagrammatically the features recognised in the field, they do show the extreme range of variability present in a turbidite unit as it is traced from proximal to distal areas of deposition. The ABC index of Walker (1967) reveals 'proximality' in terms of the a b and c divisions deposited in the upper flow regime (a-b) and the upper part of the lower flow regime (c). In the present data

the a, b and c divisions are lumped together and thus the present data represents relations between deposits due to (1) upper flow regime + upper part of lower flow regime, (2) lower part of lower flow regime, and (3) Pelagic deposition.

One difficulty experienced by the author, even in examining cut turbidites in the laboratory, is to be certain of the limits of the a, b, c, d and e divisions. This is particularly difficult in the case of junctions a-b and b-c where gradations from one to the other occur, and in the present silica-cemented protoquartzites lamination is often difficult to observe due to lack of contrasting material suitable for making lamination easily visible. Inevitably the personal observer factor is going to be important in such records of turbidite sequences and when more statistical work has been published on turbidites this may become apparent. The main point where the error is likely to occur is in the recording of beds without c divisions and when b and d divisions cannot be distinguished.

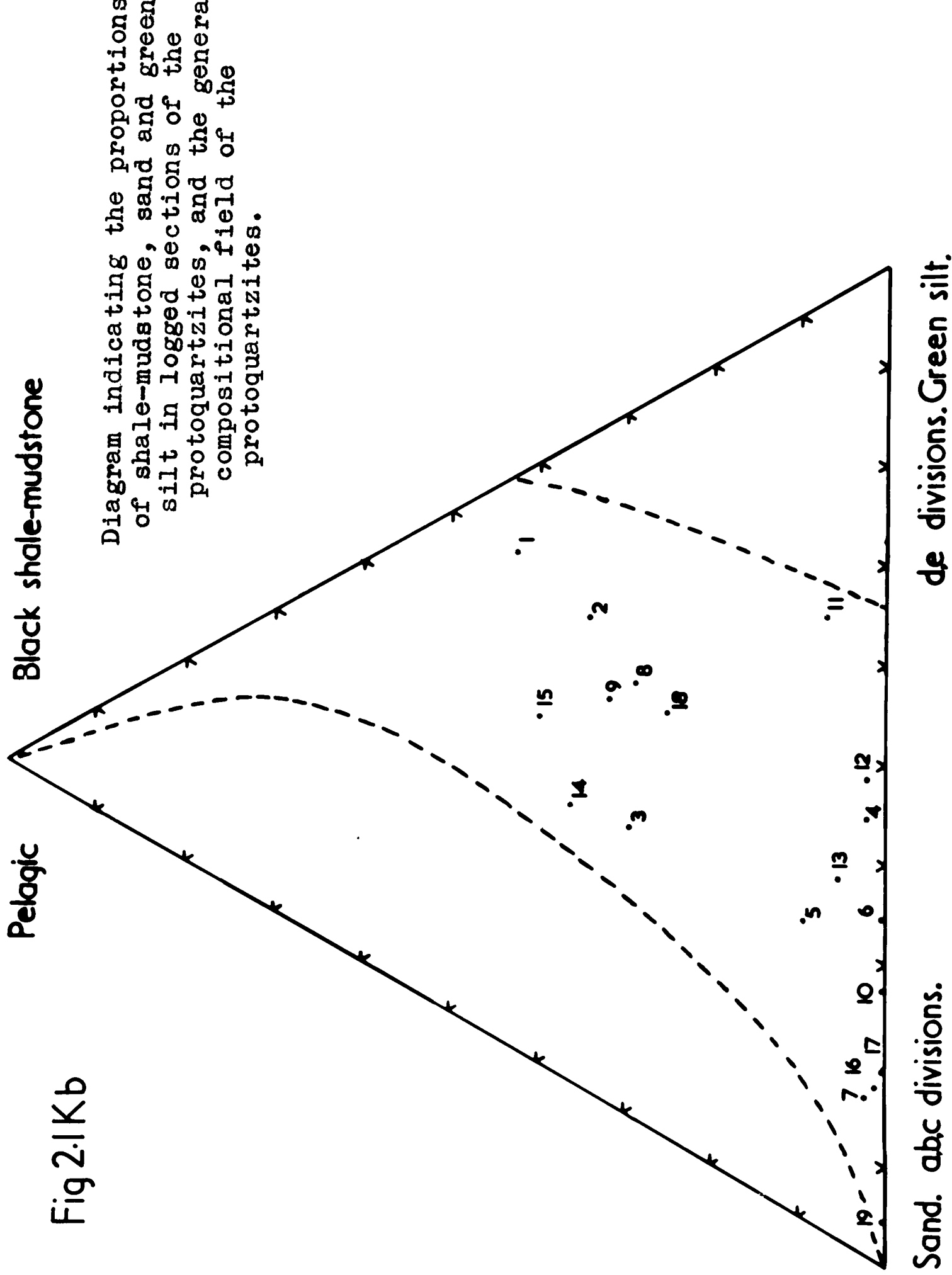


Fig 2.1Kc

RELATION BETWEEN AVERAGE SAND BED THICKNESS (a-c divisions) AND % SAND
IN LOGGED SECTIONS OF PROTOQUARTZITIC TURBIDITES

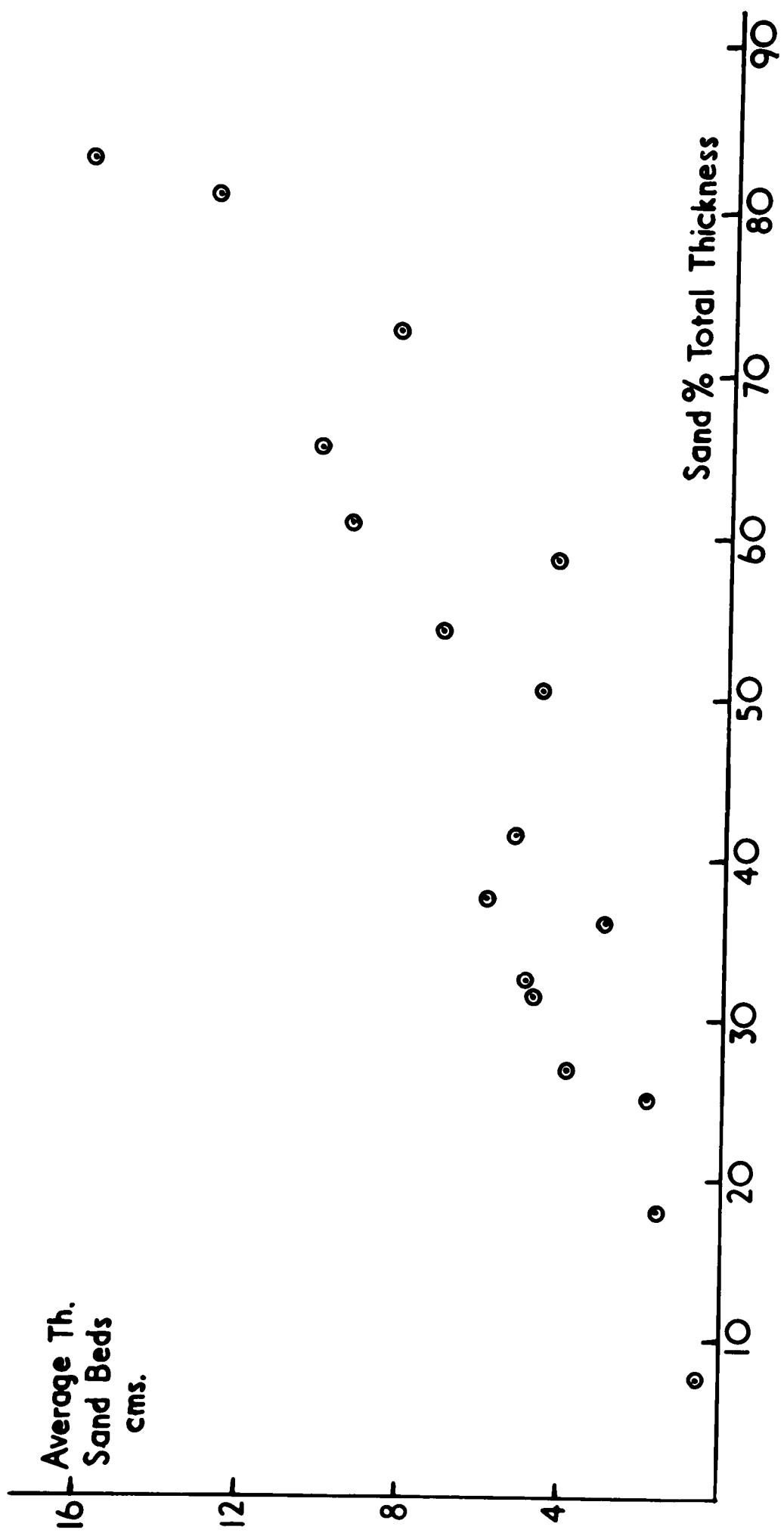


Fig 2·IKd

RELATION BETWEEN TOTAL PERCENTAGE AND AVERAGE
BED THICKNESS FOR GREEN SILT AND SHALE-MUDSTONE

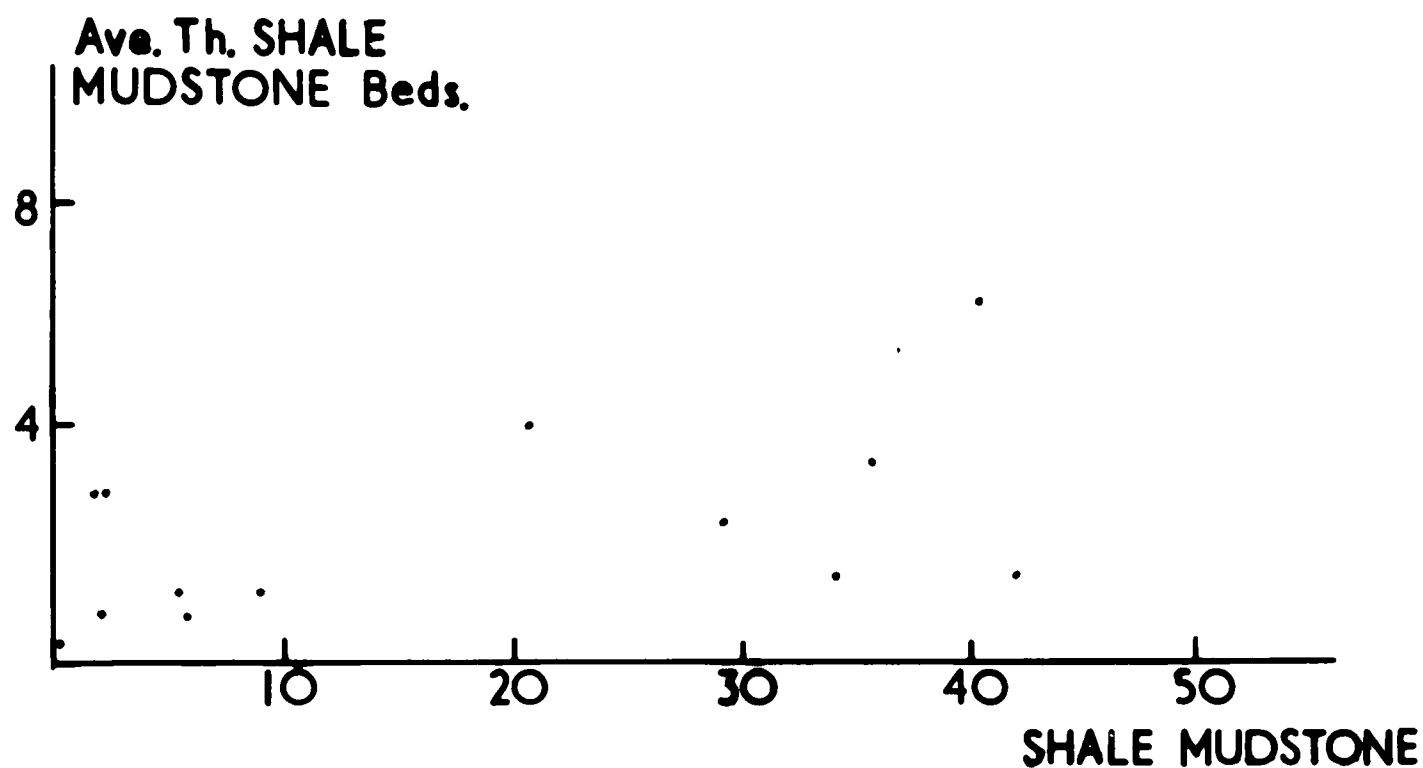


Fig2·IKe

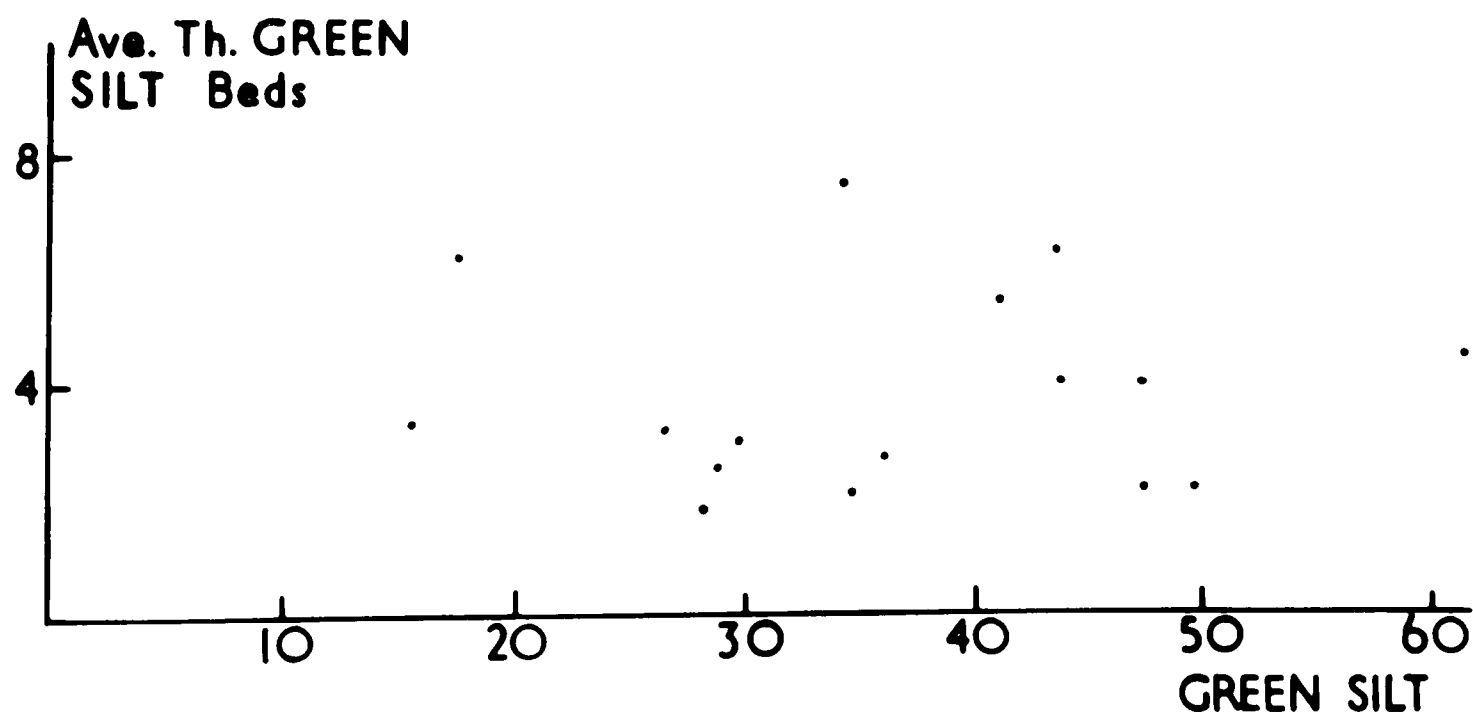
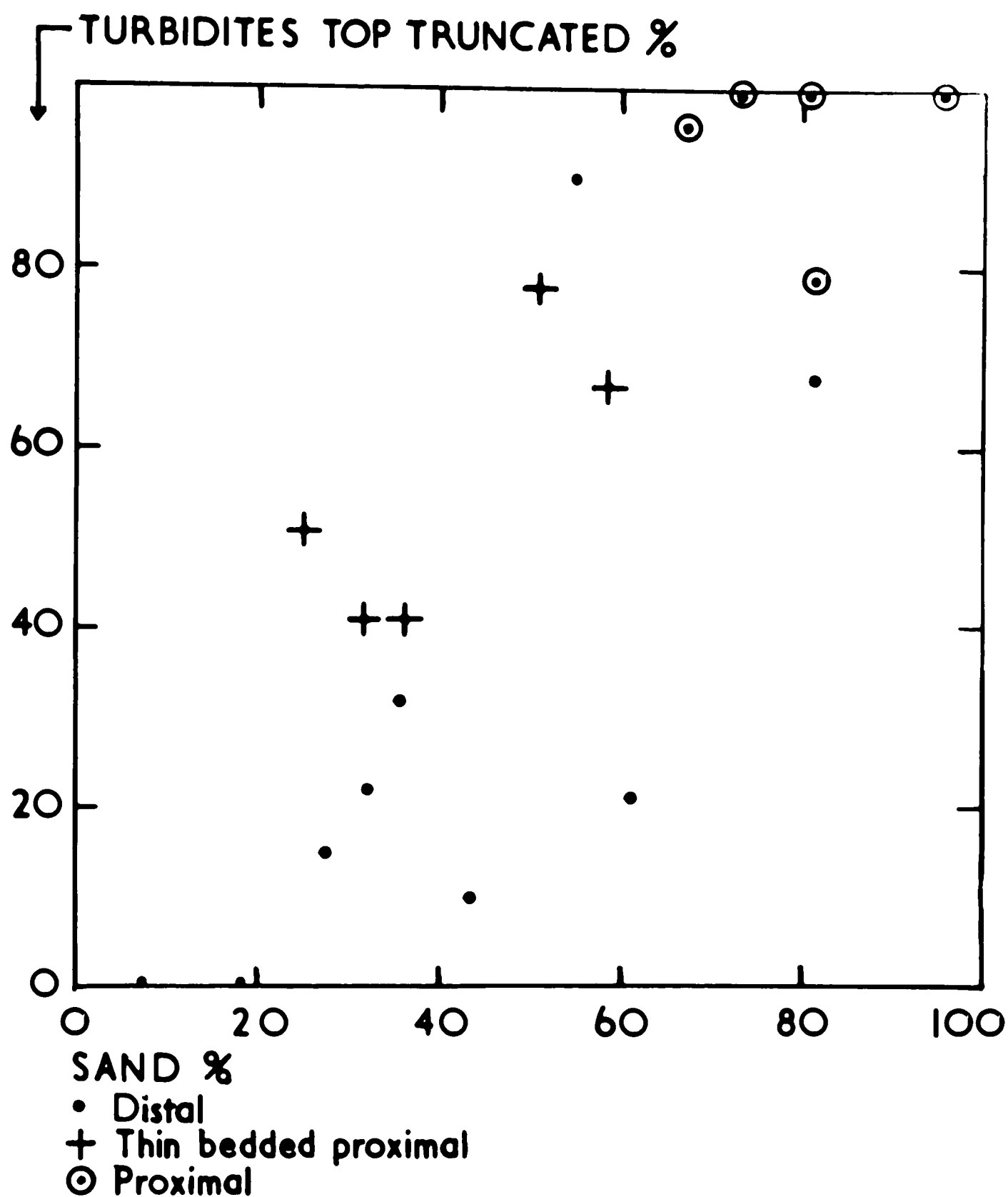


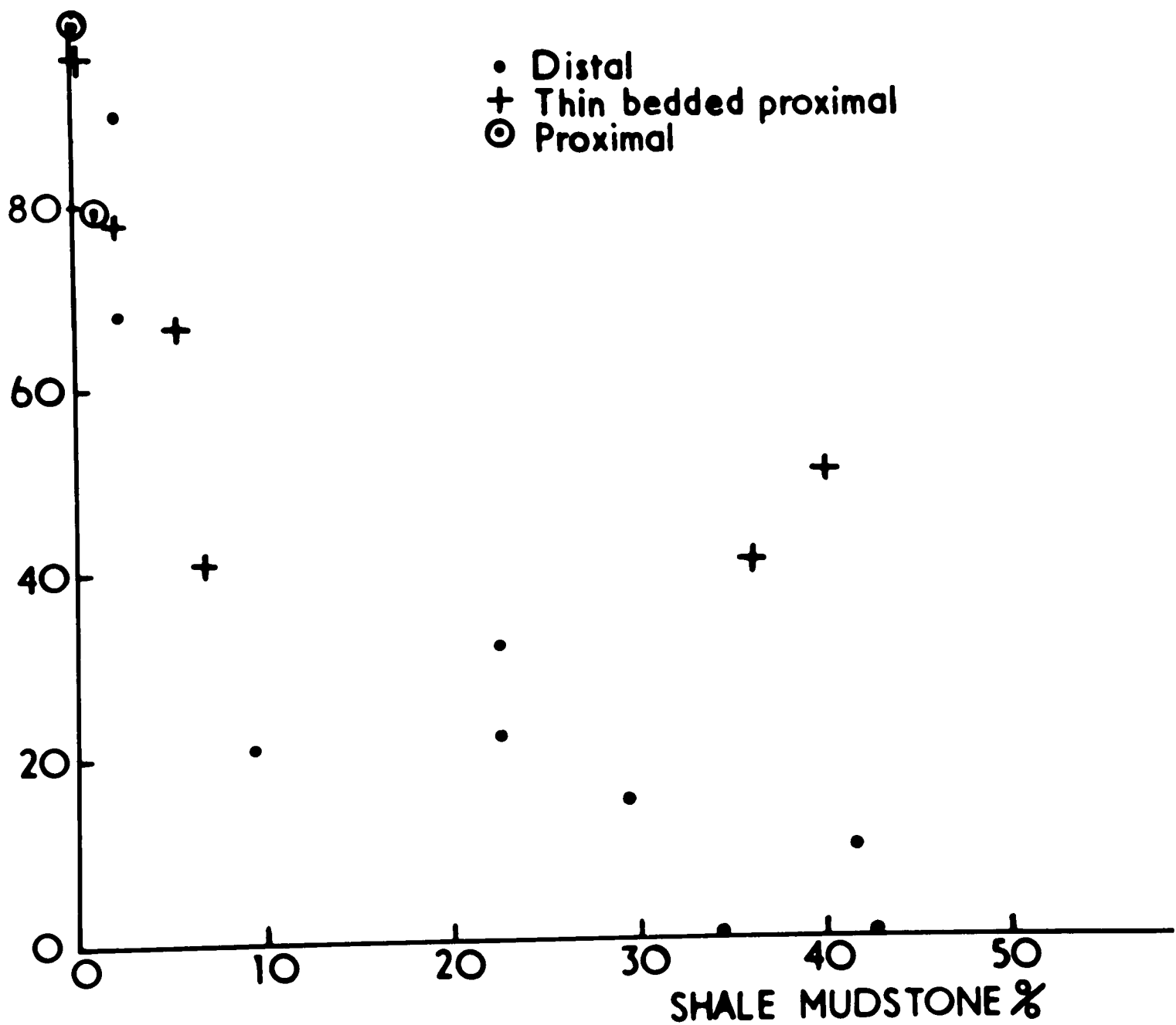
Fig 2.1Kf



Relation between percentage of turbidite beds top truncated and percentage sand forming a, b and c divisions of the turbidites in logged section of the protoquartzites.

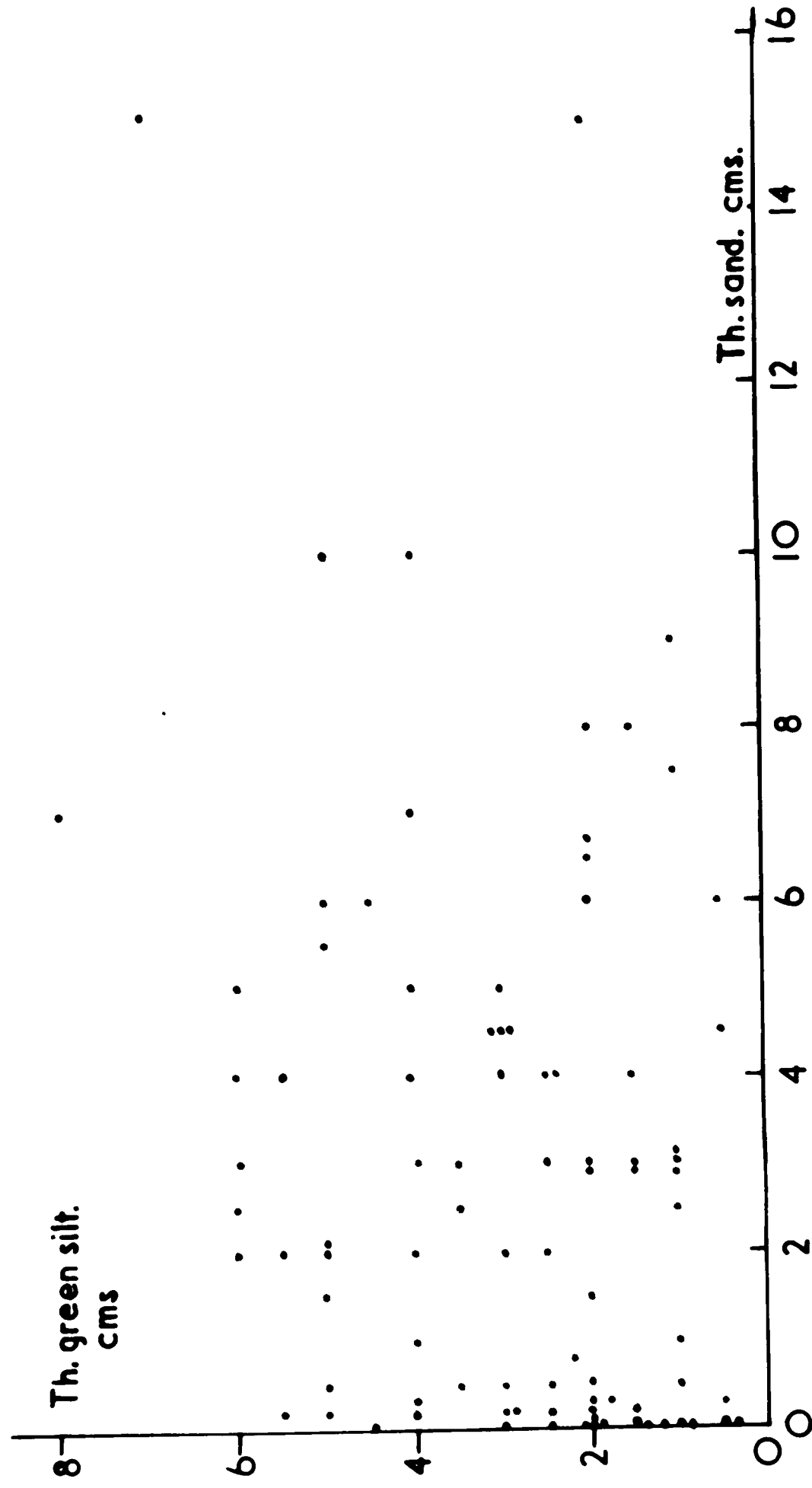
Fig 2.1 Kg

TURBIDITES TOP TRUNCATED %



Relation between percentage of turbidite beds top truncated and the percentage of shale-mudstone of pelagic origin in logged sections of the protoquartzites.

Fig 2.1Kh
 RELATION BETWEEN THICKNESSES OF GREEN SILTY TOP (d-e divisions) AND
 SANDY PART (a-c divisions) OF DISTAL TURBIDITES WHICH ARE NOT TOP
 TRUNCATED



2.2

LITHOLOGIES INTERBEDDED WITH THE PROTOQUARTZITES.Shale-Mudstone.

Between the protoquartzitic turbidite beds the dominant lithology in the area is black shale-mudstone (Fig.2.2a). It is often extremely fine grained and does not contain visible detrital grains. It does not have the perfect fissility of shale nor the typical fracture of mudstone but shows features akin to both, hence the term shale-mudstone.

In thin section little information can be gained. Sections reveal occasional quartz grains of irregular shape and variable size but rarely larger than silt grade (.0625mm.). More abundant than quartz grains are flakes of mica generally not exceeding .05mm. but occasionally .1mm. in length and oriented parallel to the bedding — thus giving the rock its weak fissility. The flakes are often twisted and 'frayed' at the ends due to alteration.

Carbonaceous material is abundant and framboidal pyrite is also present. The matrix in which these constituents are set is too fine grained for resolution using a petrological microscope (see Fig.2.2a), but X-ray diffraction work on whole-rock samples reveals something of the clay mineralogy.

A typical diffractometer trace obtained from a dark shale-mudstone sample is shown in Fig.2.2c. It reveals that the mineralogy is essentially the same as that of the d-e divisions of the protoquartzitic turbidites. Chlorite is present giving a peak at 14\AA . A broad peak from 10\AA up to 11\AA or 12\AA indicates 10\AA mica, probably detrital mica or 'illite' being altered to a mixed layer structure, thought to be a mica/montmorillonite. The interpretation of this peak has been discussed in section 1.4. Kaolinite is also present in the shale-mudstones giving a peak at 7.1\AA . The recognition of both chlorite and kaolinite depends on resolution of the 004 and 002 peaks respectively which was possible in this case. Quartz also shows on the trace.

The shale-mudstone thus contains the same clay mineral assemblage as that in the topmost part of the turbidites and it is likely that both are derived from roughly the same source area. The clay minerals, notably the chlorite and mica are less well crystallized in the shale-mudstones than in the turbidites possibly due to their finer grain size (Compare Figs.2.2a-b).

Areal variation in the basinal shale-mudstones can be detected within the area. In exposures showing distal turbidite developments the shale-mudstone is always black in colour so enabling the green turbidite bands to be clearly distinguished. As the more proximal areas are approached the proportion of black shale-mudstone becomes

less and it is replaced by a slightly more silty green mudstone. This deposit can be seen in the Hurdlow area at Locs. 009 and 003 where it occurs in the C unit of the protoquartzites.

In the proximal developments, black shale-mudstone is rare and what little sediment there is between the proximal turbidites is usually green and silty. True green and some black shale-mudstone is however present in the thin bedded proximal sequences.

It is probable that the thinning of Units A, B and C away from the source area reflects not only the dying out of the turbidites but also a reduction in thickness of the normal bottom sediment. It is thought that the shale-mudstone was deposited from suspension and that the source of the material was the same as that for the protoquartzites.

The material could represent the suspended load of rivers draining into the area from a land mass to the south, this material being distributed over the whole basin but spread most thickly near its source. The protoquartzites are interpreted as being derived from a deltaic area and thus the same river system was probably the source of both lithologies. It is of course possible that fine material was brought into the area from other directions but this supply was probably only of minor importance.

Bands of siderite nodules which have the form of discoidal masses with an elliptical cross section are of frequent occurrence within the black shale-mudstone between the protoquartzites. They are more frequent in the distal areas, but they do occur even in the proximal developments. The significance of the siderite in the shale-mudstones and in the protoquartzites is discussed in Part 5 of this thesis.

Pebbly Sandy Mudstones.

In the sections of proximal turbidites at Shirkley, Endon and Sprink a few beds occur that consist of grey-green, red or purple mudstone with an admixture of quartz of all grain sizes up to pebbles (Fig.2.2d). The beds are all poorly exposed and little positive evidence can be gained from them. At all localities, scattered round quartz pebbles are present and the beds usually contain crude laminae of quartz grains (Fig.2.2e). Plant fragments are frequent and at Shirkley seeds of plants were found in one bed. Mud flakes are also frequent in the beds. The clay in which the quartz grains are set is very rich in kaolinite with mica and chlorite subordinate. Kaolinite is approximately twice as abundant in these beds compared with the black basinal shale-mudstone.

The origin of these beds is obscure since so little of them is exposed. They do not appear to be slump

deposits since lamination of sand grains is present and shale pellets are well rounded, and there is no evidence of disturbance within the beds. If the beds are not slumped then the problem remains of depositing all grain sizes from clay to pebbles in a single bed. The kaolinite rich nature of the clay fraction probably indicates that the deposit was formed near the point of discharge of the land derived material into the basin area. In the marine environment kaolinite is altered to illite and this cannot have happened to any great extent in these beds. If the deposit was formed near the point of discharge of a river into the basin area the variable deposit can possibly be explained in terms of material being dropped from floating debris brought down by the river.

Current action has certainly affected the deposit as can be seen from the lamination in Figs.2.2d-e, but it seems unlikely that the same current could be responsible for the presence of the pebble in Fig.2.2d. The laminae produced are irregular and lack sharp contacts above and below. The laminae are also discontinuous laterally. Winnowing of the surface material by a variable current could cause this type of lamination.

The beds described here are given the name pebbly-sandy-mudstones to stress their unsorted nature. They are interpreted as forming near the point of influx of

land derived material into the basin. They were possibly deposited during periods of river flood and heavy discharge of material from the land. In such conditions a river would be capable of bringing greater variety of grain sizes to the area and vegetation torn from the river banks could transport the coarse sand and pebbles trapped in its roots. The rapid slowing down of the currents on entering a larger body of water could account for the deposition of such a variety of material in one bed.

The pebbly-sandy-mudstones are always found interbedded with the proximal type protoquartzites where shale-mudstone is virtually absent. They are not associated with the quiet conditions of shale-mudstone deposition, and it is probable that there was continuous slight current activity over the most proximal areas which did not allow the fine shale-mudstone to be deposited. This would account for the small proportion of material between successive turbidites.

Green Silt with Ripples.

Interbedded with the proximal turbidites are beds of green silt similar to that forming the d-e divisions of turbidites in the distal parts of the basin. Within these green silt beds isolated ripples of sand may occur with an amplitude of up to 1.5cm. and wavelength of 10cm. It is not possible to be certain that these ripples were

not a product of the currents from which the proximal turbidites were deposited, but in view of the general lack of ripples in proximal turbidites it is likely that they are due to some other mechanism.

Normal currents winnowing the fine material from the green silt could be responsible for forming the ripples. The absence of fine shale-mudstone in proximal turbidite sequences could also be due to such currents.

Distribution of deposits interbedded with the protoquartzitic turbidites.

The following types of sediment have been recognised as not being deposited from turbidity currents.

1. Pebbly-sandy-mudstones.
2. Green silt with isolated ripples.
3. Green silty mudstone.
4. Black shale-mudstone.

Types 1 and 2 are confined to the sequences of proximal turbidites (see log) and are interpreted as being near source deposits of land derived material mainly from suspension during time of heavy sediment discharge of rivers. They have been modified by currents capable of producing weak parallel lamination and ripple lamination. No evidence of currents stronger than those required for rippling of sand has been found. Grains ranging from pebbles to sand in size were probably

introduced by floating vegetation, a theory supported by the abundance of plant material in the area.

Sediment types 3 and 4 are gradational one to the other, and are more characteristic of distal environments. The black shale-mudstone represents the finest material carried in suspension and is the typical deposit of distal areas close to the eastern edge of the basin. Green silty mudstones are more characteristic of environments intermediate between proximal and distal such as those at Thorncliff and Hurdlow.

The main distribution feature of the deposits interbedded with the protoquartzites is the reduction in maximum grain size away from the area interpreted as proximal on turbidite evidence. In proximal regions laminations and ripples are interpreted as representing weak normal currents.

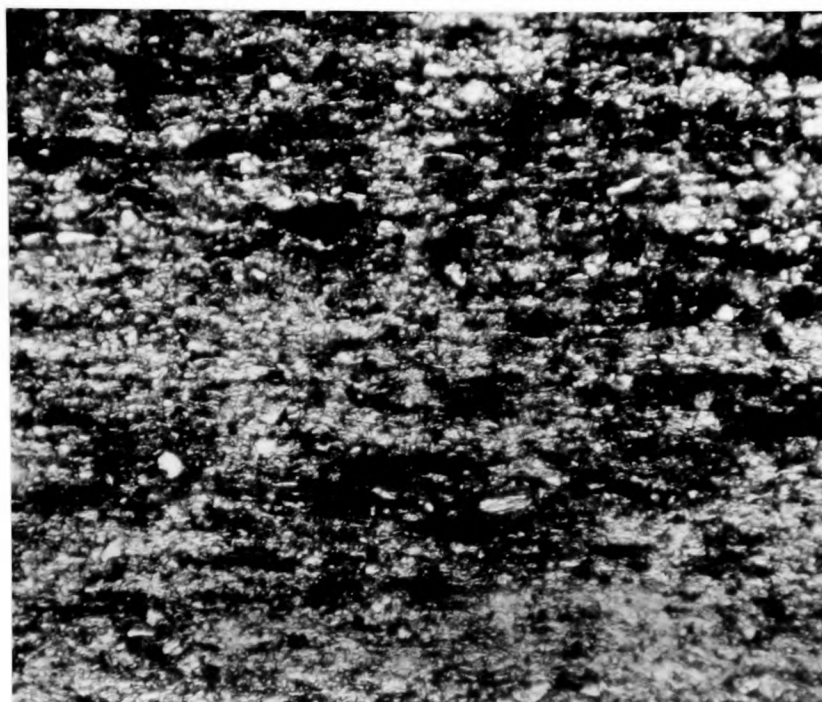


Fig. 2.2a

Thin section of black shale-mudstone from between protoquartzite beds. Slide 102.5a. Unit A. Loc.102.

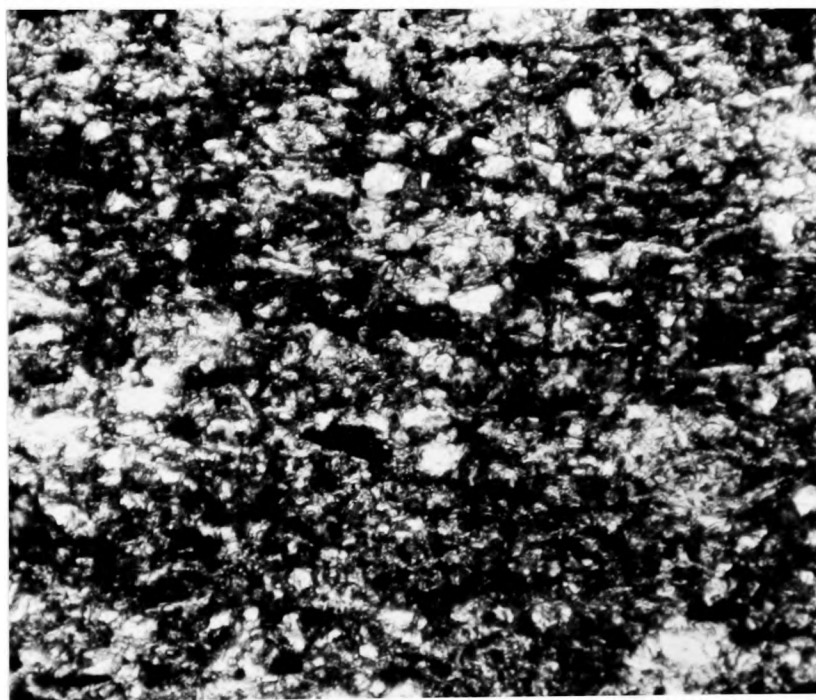


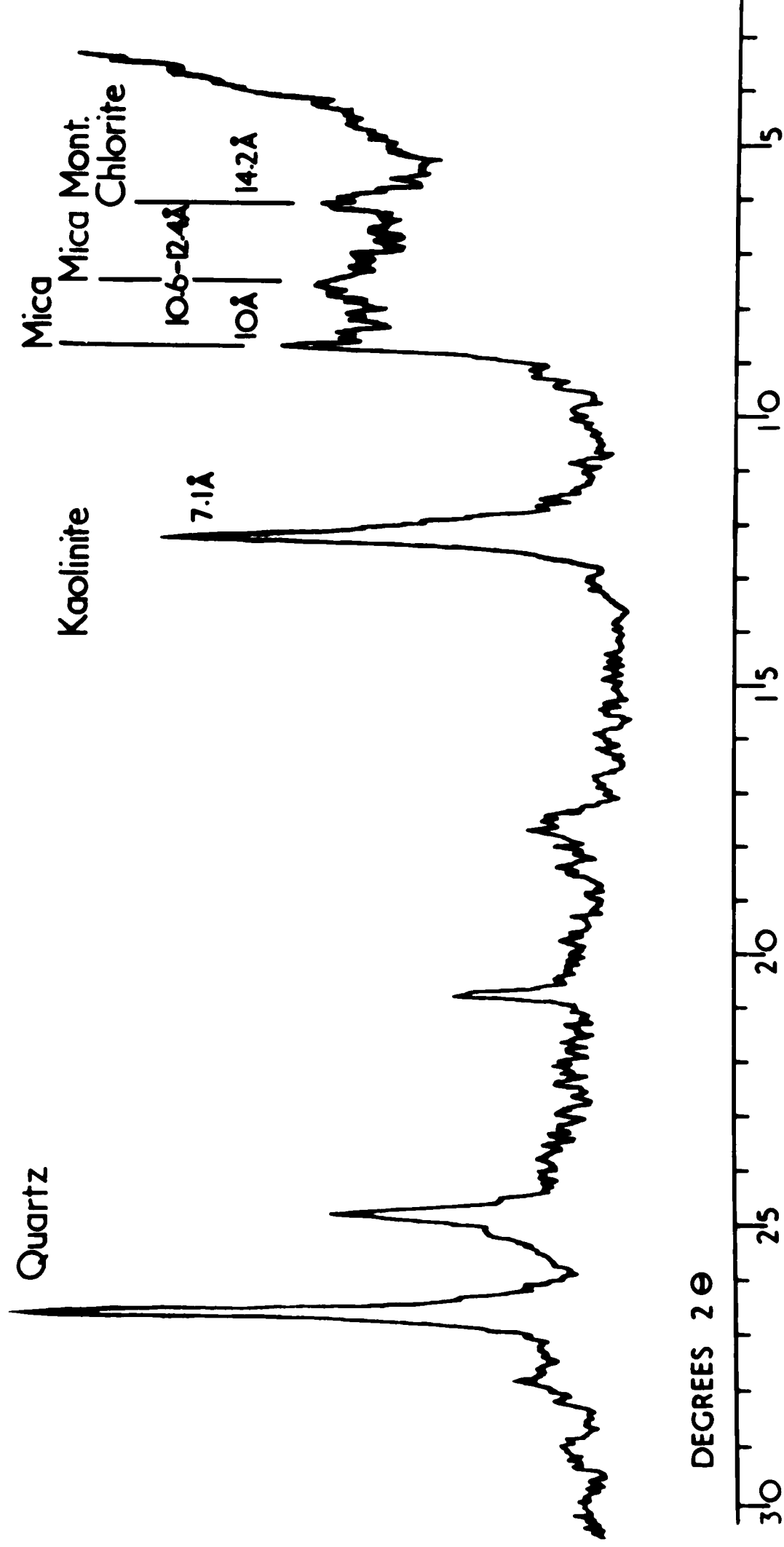
Fig. 2.2b

Thin section of green silty mudstone which forms the e division of the turbidites, and the normal basin sediment in the proximal areas. Same scale as Fig.2.2a for comparison. Slide 102.4. Unit A. Loc.102.

Fig 2:2c

Shale Mudstone

Spec. 32. sample untreated.



Diffractometer trace of typical shale-mudstone with important reflections indicated.



Fig. 2.2d

Pebbly sandy mudstone showing isolated quartz pebble set in weakly laminated sandy mudstone. Unit C Upper Shirkley.

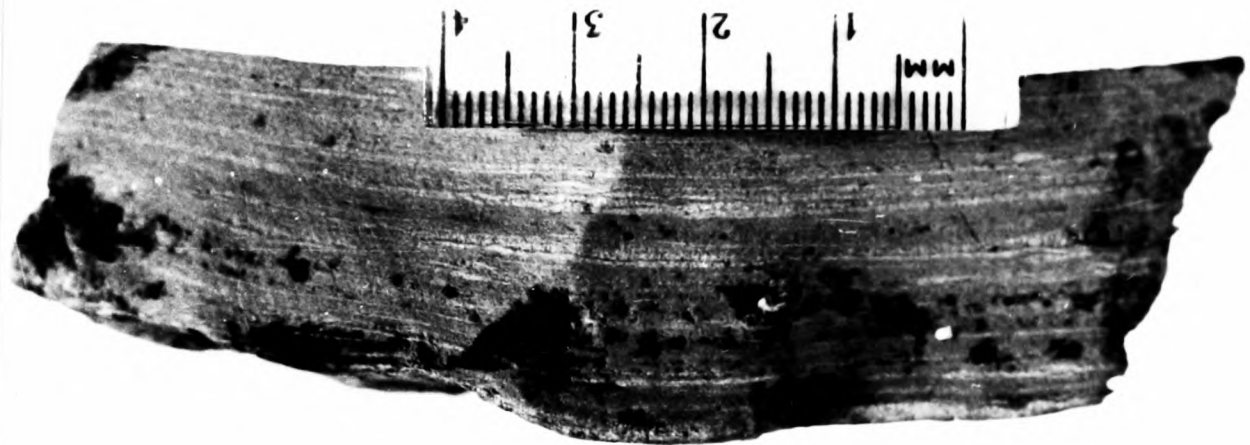


Fig. 2.2e

Weak parallel lamination in sandy mudstone. Unit C Upper Shirkley.

2.3

UNIT OF SHALE-MUDSTONE WITH SIDERITE.

It has already been stressed that the deposition of protoquartzitic turbidites is strongly associated with developments of black shale-mudstone with siderite beds and concretions. It can be demonstrated that the lateral equivalent of units A, B and C is shale-mudstone with siderite nodules where the turbidites are absent.

Beneath the E. ferrimontanum-E. erinense faunal band there is a unit of shale-mudstone with siderite nodules which is of remarkably constant development all over the north Staffordshire basin area and the Edale area. No turbidites or other coarser grained rocks are found within the unit although the diagenetic conditions are identical with those which prevailed during deposition of the protoquartzites. This indicates that the introduction of the coarse grained material in units A, B and C is not directly responsible for the different chemical conditions of these units. It is probable that the introduction of the land derived protoquartzitic turbidites in one area was associated with fine grained deposition from the same source over a much larger area than that covered by the turbidites. This is shown by the correspondence of siderite deposition in Edale (Part 3.2) with protoquartzite deposition in north Staffordshire. The diagenetic

significance of the siderite is considered in Part 5 of this thesis.

The persistent unit of shale-mudstone with siderite below the E2a2 faunal band is probably the equivalent of a protoquartzitic turbidite development somewhere outside the present area. A possible area for such deposition is in the Windmerpool Gulf south of the Derbyshire massif (Falcon and Kent 1960) where turbidites are also present in E2a on borehole evidence (Ramsbottom pers. comm.) but no published evidence is yet available on their exact stratigraphic position.

In the shales of this unit kaolinite is abundant and both well crystallized 10Å mica and chlorite are present together with some mixed layer material. This is an identical assemblage to that found in association with the protoquartzites and strengthens the view that this unit of shale-mudstone with siderite nodules was derived from the same environment as the protoquartzitic material of units A, B and C.

Fig.2.1Cb.

LOG. PROXIMAL PROTOQUARTZITES.
Unit B. Loc.072.

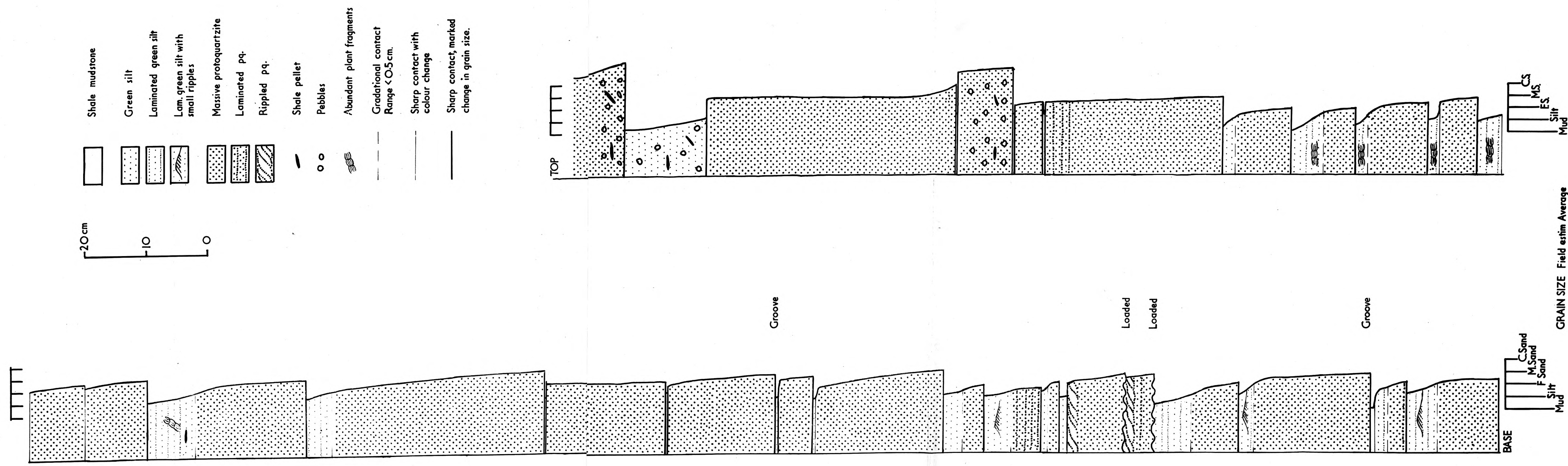


Fig.2-1Cc.

LOG. DISTAL PROTOQUARTZITES.

Unit A. Loc. O12.

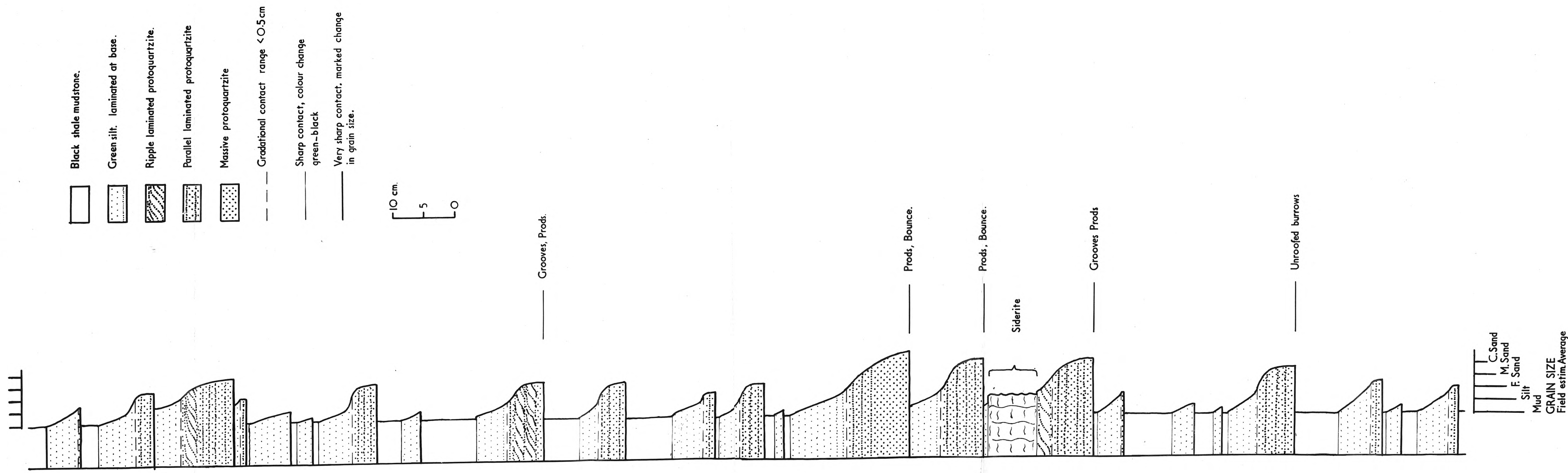
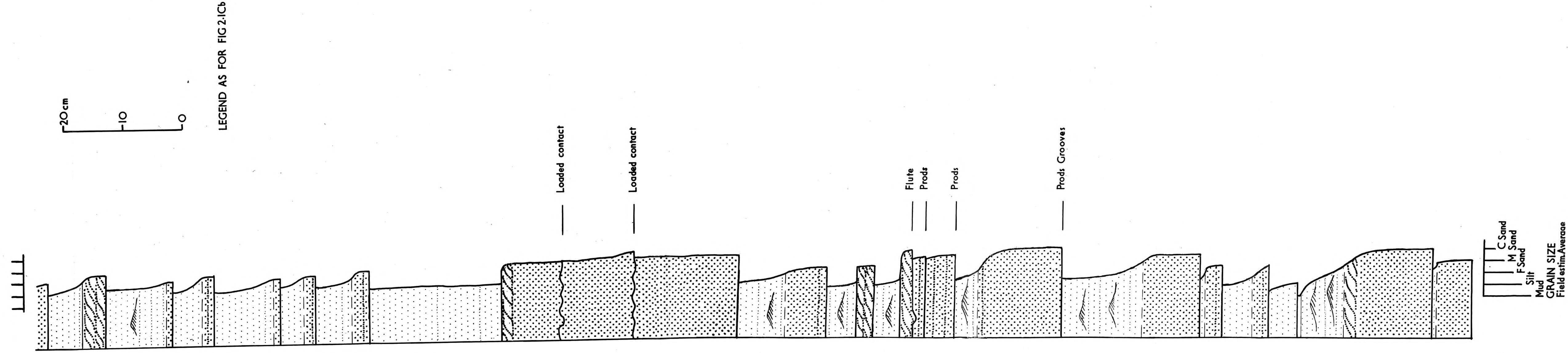


Fig 2-1Cd.

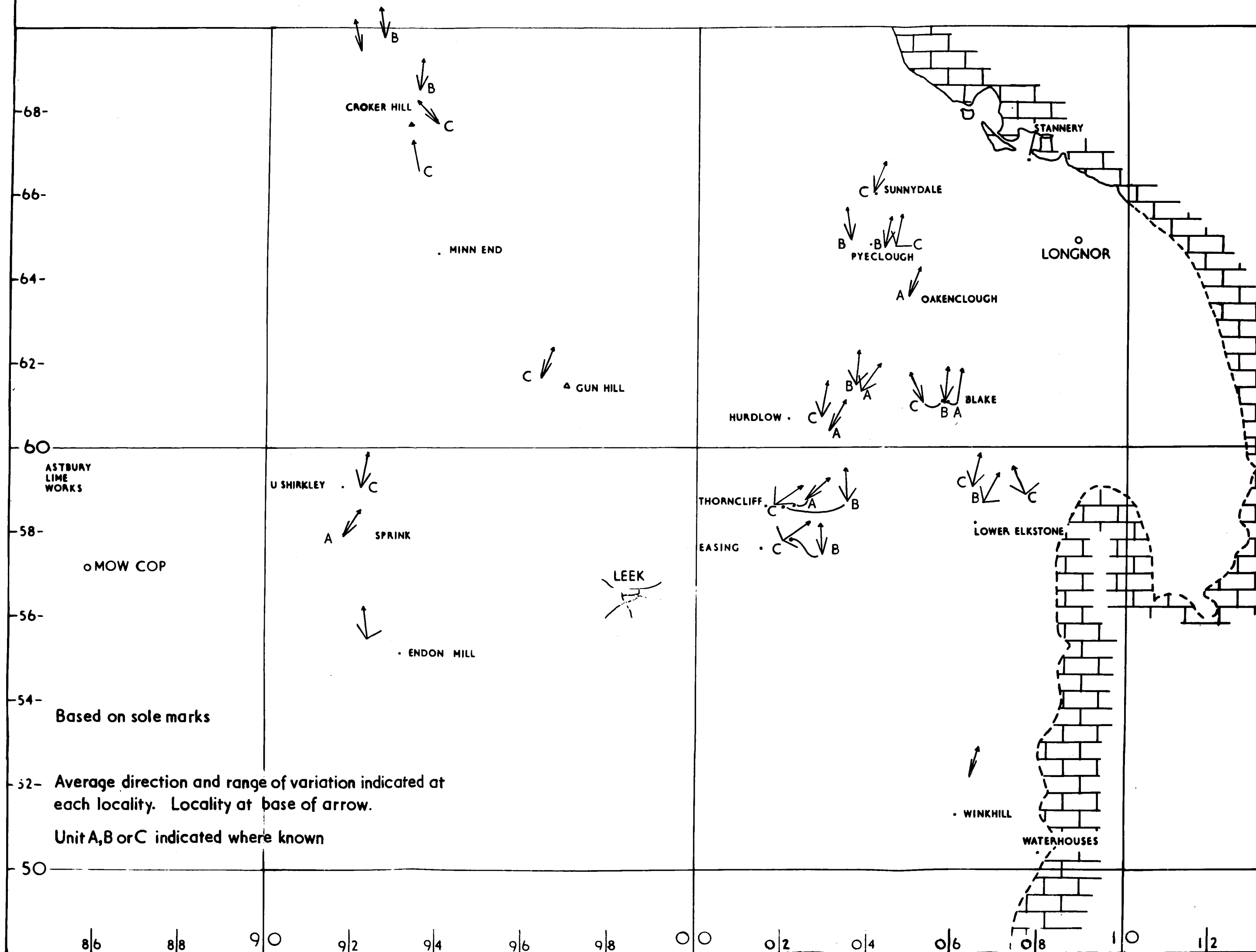
LOG. THIN BEDDED PROXIMAL PROTOQUARTZITES.
Unit C. Loc.O51.



CURRENT DIRECTIONS

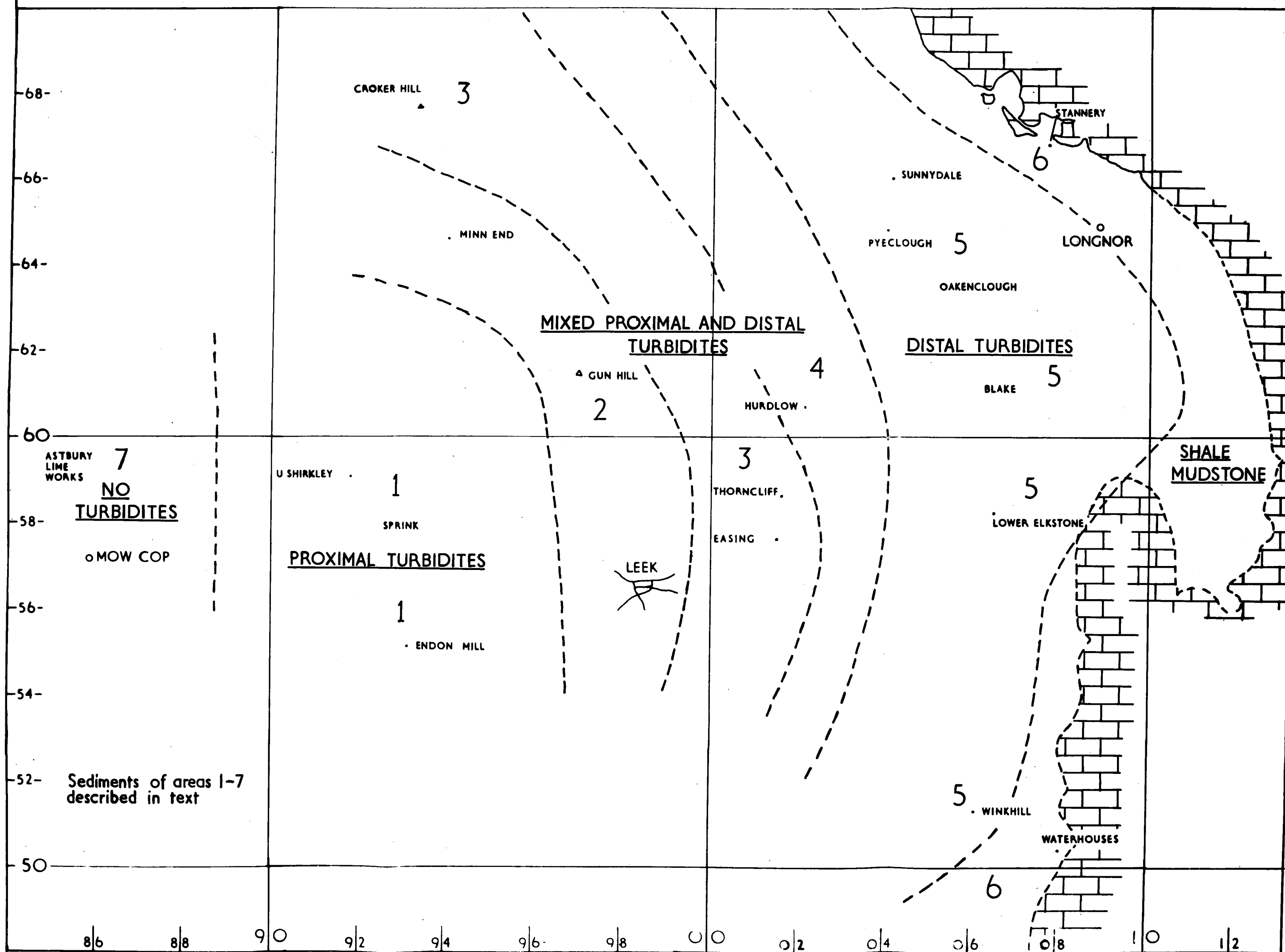
PROTOQUARTZITIC TURBIDITES

Fig 2.1 Ja



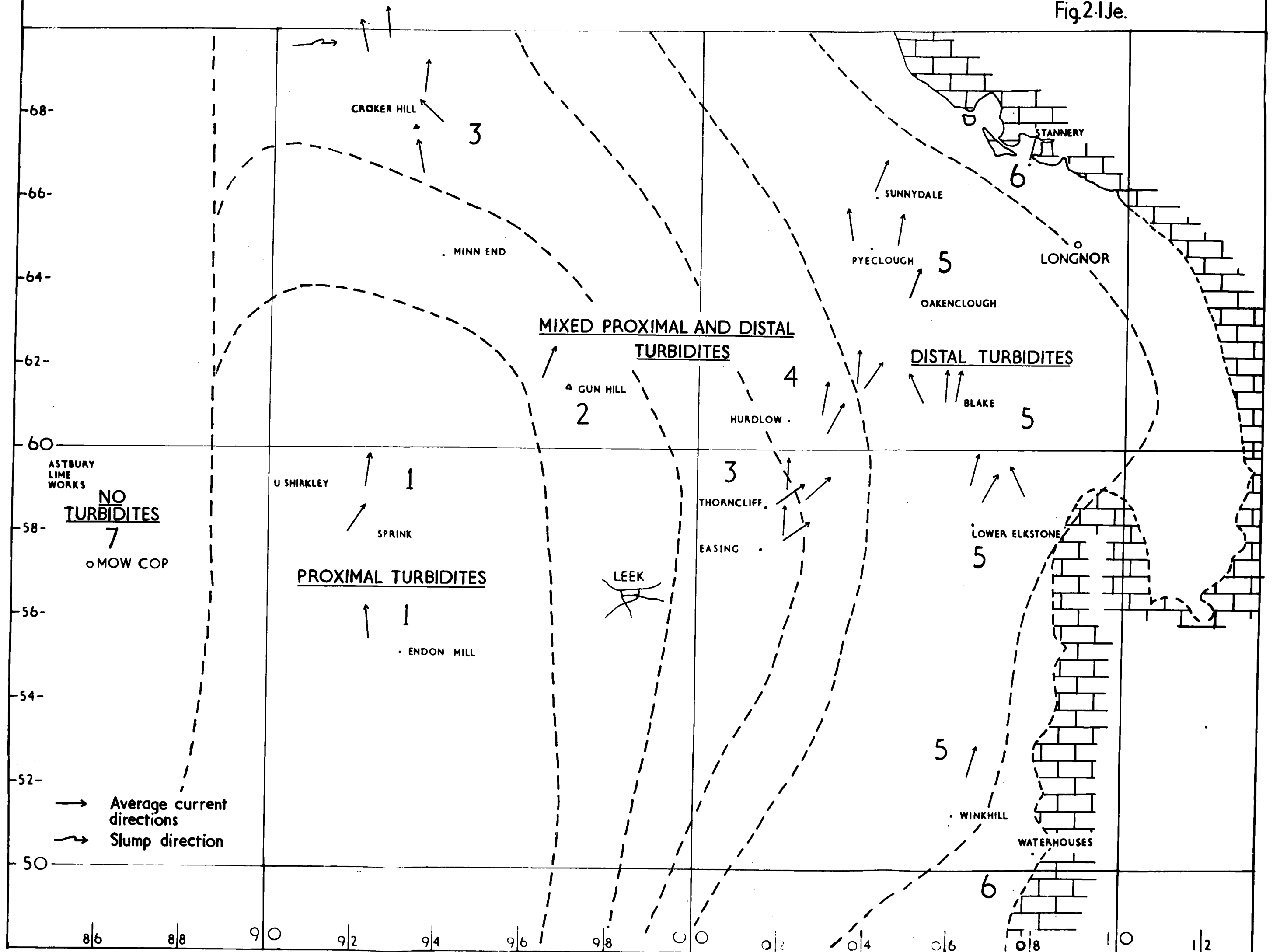
DISTRIBUTION OF AREAS OF PROXIMAL AND DISTAL TURBIDITE DEPOSITION

Fig. 2.1 Jc.



AREAS OF PROXIMAL AND DISTAL TURBIDITE DEPOSITION RELATED TO CURRENT DIRECTIONS.

Fig.2.1Je.



SECTION 2.4

THE CALCAREOUS SILTSTONES.

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SECTION 2.4

THE CALCAREOUS SILTSTONES.

2.4A

BRIEF DIAGNOSIS.

Parallel sided beds up to 20cm. thick often graded and usually laminated, sometimes with ripple lamination, sole markings absent. Beds contain transported marine organic debris. Cemented and replaced by calcite ankerite and dolomite but not siderite. Interbedded with shale and shale-mudstone with P. corrugata.

2.4B

POSITION IN SUCCESSION.

Three units of the calcareous siltstones are present in the succession studied. The lowest is Unit α which is always present below Unit A of the protoquartzitic turbidites (Fig. 1.1a). Unit β is immediately above the C. malhamense (E1c) faunal band and Unit δ immediately above the E. grassingtonense (E2a1) faunal band. The top part of the E. erinense - E. ferrimontanum (E2a2) band could be likened to a fourth very weak development of calcareous siltstones and is discussed in section 2.6 along with other marine band sediments as it contains goniatites.

The lowest unit present is the equivalent of the Gun Hill Siltstones Formation of Holdsworth (thesis 1963

p. 170) and of 'The Crowstones' of Hudson (in Hudson and Cotton 1945 p.321) which occurred from 280-830 feet in the Gun Hill Borehole. Units β and γ occur within the 'Thorncliff Sandstone Member' of Holdsworth (thesis 1963 p. 170).

The calcareous siltstones described here correspond to Lithofacies 1 of Holdsworth (thesis 1963 p.187). He divided the lithofacies into two sublithofacies, one containing beds with only weak lamination, individual pale laminae not exceeding 3mm., and the other of units with roughly half the beds showing pale laminae of 3mm. or more. The author prefers to consider the units of calcareous siltstone together and while recognising that both type of beds exist has not noticed beds with coarse lamination to be predominant except over short measured intervals of about a metre.

Areal Variation.

In general there are no significant differences between units α , β and γ , and over the main part of the basin they exhibit very little variation. The following descriptions apply to most outcrops and the few departures from the normal type will be discussed when the areal significance of the calcareous siltstones is considered.

2.4C

SEDIMENTARY STRUCTURES.

A. Structures occurring as sole markings.

Sole structures are virtually absent from the calcareous siltstones. No tool or scour markings or moulds have been seen by the author in the field. Holdsworth (thesis 1963 P1.19) figures a protruberance on the base of a bed which may represent an isolated flute mould, and mentions (Ibid p.207) an isolated example of a delicate groove mould. No load structures are present on the bases of beds, and there is no evidence of any burrowing activity on or within the beds.

The only possible example of a sole structure seen was doubtful grain striation on the base of a bed at Loc.109 Pyecrough. The trend of the striations was 180° but as this is also parallel to the axes of the folds in the area the weak striations seen may be tectonic in origin.

B. Internal Structures. Figs. 2.4Ca - r.

Internally the beds are usually laminated, parallel lamination being much more frequent than ripple lamination. In beds less than 10cm. thick the dominant lamination consists of pale quartz rich laminae up to about 5mm. in thickness but usually less than 3mm. set in dark siltstone. The dark coloured, matrix rich, material is always the more abundant. In the thicker beds (up to 20cm.) the laminae near the base are coarser

and pale coloured, matrix-free material may be dominant over the dark siltstone. The bases of the thicker beds often contain material of fine or medium sand grade and occasionally a few grains of coarse sand are present near the base of the bed. When these coarse, pale laminae are present, the bed may show ripple lamination. The ripples consist of pale, quartz rich material from which the dark matrix material was presumably removed by sorting. The sequence of structures seen within a thicker bed from base to top is: coarse parallel lamination - ripple lamination - fine parallel lamination - dark silty graded mudstone. The ripple interval is very variable in occurrence and this important feature is considered below.

Coarse basal laminations. (see Figs. 2.4Ca,o,p,q,r).

The basal coarser laminae are usually few in number, one or two being usual. They range up to a centimetre in thickness, but at this thickness finer laminae can often be seen within the larger pale layer (Fig. 2.4C a,o,r.); This finer lamination is not always perfectly parallel to the top or base of the major laminae and may represent very low amplitude ripples. The coarse laminae contain small shale pellets similar to the shale that separates the beds, and also contain material visibly coarser than that found in the pale laminae in the upper part of the bed. Separating the pale laminae are dark, clay rich

laminae containing scattered quartz grains of similar size to those present in adjacent pale, quartz rich laminae (Holdsworth (thesis 1963)).

The beds illustrated with basal parallel laminae greater than 4mm. thick fall into two groups. In group (1) Figs. 2.4C o,q,r, the basal pale laminae are sharply divided from the intervening dark laminae, and the laminae are followed by 2-5cm. of dark siltstone containing only weak pale parallel laminae. Group (2), Figs. 2.4Ca,b,c,p. have the basal pale laminae only poorly divided from the intervening dark laminae, there being a general mixture of the pale and dark material. The parallel lamination in this group is followed by ripple lamination in all cases, sometimes associated with convolutions. The ripple lamination in these cases comprises 5cm. or more of the bed and individual ripples have an amplitude greater than 1cm. Dark clay material has generally been removed from the ripple zone so it forms a pale portion of the bed. Beds of the second type with parallel lamination followed by ripples are scarce in the calcareous siltstones.

Ripple Lamination.

The ripple lamination observed in the calcareous siltstones is variable in nature. As already mentioned ripples are usually fairly well sorted and contain little of the dark, clay-rich siltstone. Ripples that follow

initial parallel lamination have amplitudes of up to 2 or 3cm. The ripple amplitude is however variable and it appears to increase gradually upwards from parallel lamination (Fig.2.4Cd) and also to die out upwards into parallel lamination, (Figs. 2.4Ca,b,d,m,n.). The upward disappearance of ripples with incoming of parallel lamination is significant with respect to the coarser parallel lamination described. It appears that the coarser parallel laminae of group (1), which are sharply divided from intervening dark siltstone horizons, and in which very weak inclined lamination is observed, (Figs. 2.4Co,r.), represent the final stages of ripples dying out to parallel lamination (compare Fig.2.4Cd top with base of 2.4Co for example).

Another type of ripple lamination seen in the calcareous siltstones consists of isolated ripples enclosed in dark siltstone. The ripples may be very small ($a=2\text{mm}$, $\lambda=10\text{mm}$ Fig.2.4Cr) or larger ($a=1\text{cm}$, $\lambda=7-10\text{cm}$, Figs. 2.4C h,i,k,l). The ripples are frequently isolated and have loaded bases with flame structures extending up from the dark siltstone beneath. Usually only an isolated row of ripples is present but occasionally enough are present to form a continuous bed of rippled sand (Fig. 2.4Ch). The ripples have sharp tops against dark silty mudstone which usually grades to shale

in a few centimetres. Beneath the ripples weak parallel lamination is usual, this being disturbed immediately below the ripples.

In the few examples seen the ripples appear to be essentially similar to those already described. The overall shapes of the ripples could not be determined at outcrop, but sectioning reveals that they are not parallel crested and are probably linguoid in shape.

Fine parallel lamination.

The fine parallel lamination seen in Figs. 2.4Ce,q is typical of that which occurs commonly within the calcareous siltstones. The individual pale laminae are less than 3mm. thick and usually have sharp upper contacts with dark siltstone. The bases of the laminae are less sharp, due to incorporation of the pale, quartz rich material within the underlying dark siltstone by a process of micro-loading. Since the tops of the pale laminae are undisturbed, the loading took place syndepositionally.

Beds with weak parallel laminae are the ones most frequently met within the calcareous siltstone units. Often the base of the bed is marked by a pale lamina (Figs. 2.4C f,g.), but this is not always the case. The laminae are most frequent in - and are usually confined to - the lower halves of beds which consist of dark siltstone grading up into a silty mudstone at the

tops (see Holdsworth thesis 1963 p.191). Fig.2.4C f, is typical of the very weakest of the beds seen, consisting of a basal pale lamina, a few grains thick, followed by a few centimetres of silty mudstone - only slightly paler than the shale which separates the beds.

C. Distorted lamination in calcareous siltstones.

Several types of distortion have been observed and each type is typical of a particular part of a bed.

Convolute lamination - Colvolute ripple lamination.

Convolute lamination is found rarely (two examples seen) and is developed in the lower coarse laminae which are typically overlain by ripple laminae. In cases such as Fig. 2.4Cb it is gradational to convolute ripple lamination, since the sand which fills the sinking synclines is rippled while the original distortions were developed in parallel laminated material. One example of a bed was seen where anticlines in original convolute lamination were clearly truncated by later colvolute ripple lamination.

Distorted parallel lamination.

Towards the top of several beds a markedly pale zone is developed in very fine sand and coarse silt - grain size being difficult to determine due to extensive secondary replacive carbonate (Figs. 2.4C o,q,r.). This zone has thicker laminae than the fine ones discussed

and the bases of the laminae are always very irregular, having flames of clay rich siltstone of variable size extending up into the pale laminae (Figs. 2.4C q,r). Thin clay rich laminae may become totally mixed with the pale laminae, which then have irregular internal distortions as in Fig. 2.4Co . The tops of these disturbed pale laminae are usually perfectly parallel and the distortion must have taken place during deposition of the pale laminae.

Distortions within dark silty mudstone.

The top of the bed in Fig.2.4Cb shows a recumbent fold of pale material within dark silty mudstone. Similar distortions can be seen in Figs. 2.4C j,p,r. These distortions are due to the incorporation of pale material within the silty mudstone, but in these cases a planar top is not developed to the pale material, and incorporation and distortion probably occurred in part postdepositionally.

D. Grading.

Grading can be clearly seen in hand specimens of the calcareous siltstone but it is obscured by the rapid changes in sorting that occur within the beds to give pale and dark laminae. Holdsworth (thesis 1963 p.201) showed that there is an overall decrease in maximum grain size upwards in the beds. The sudden changes

in sorting that occur make the average grain size in the dark, clay rich portions of the bed less than that in the pale laminae, but the quartz grains scattered in the dark siltstone are of roughly the same size as those in adjacent pale, quartz rich laminae. At the top of the beds the dark silty mudstone frequently grades to ordinary shale without any perceptible break.

E. Ideal Sequence of Structures within Calcareous Siltstone Beds.

Fig. 2.4Cs illustrates the sequences of structures observed within the calcareous siltstone beds. The main features are as follows:-

1. The various possible base cut-out sequence which give beds starting with coarse parallel lamination, ripple lamination or fine parallel lamination.
2. The variable position of rows of isolated ripples in dark siltstone within individual beds.
3. The evidence of syn- and post-depositional distortion within the beds.

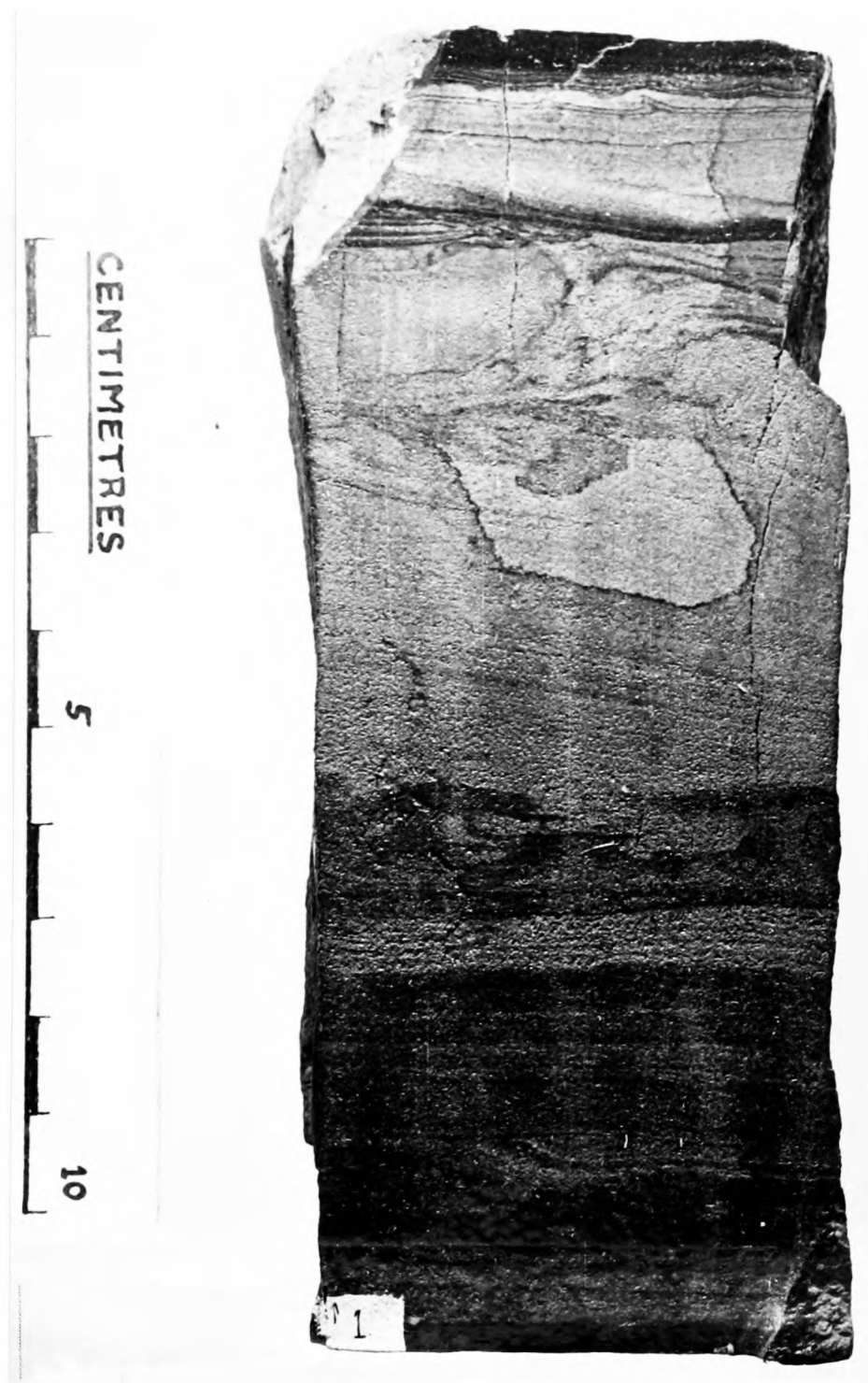


Fig. 2.4Ca

Calcareous siltstone bed showing sequence from coarse basal laminae through ripple laminae, which are partly convoluted, to fine parallel laminae and a thin dark graded siltstone top. Unit 8 . Loc.353.



Fig. 2.4Cb

Thick calcareous siltstone bed with coarse basal laminae followed by convolute lamination and ripple lamination. Shale fragments enclosed in the ripple lamination. The top half of the bed consists of fine parallel laminae and dark siltstone. Syndepositional distortions in quartzose silt occur at the top of the bed. Unit 6 . Loc.353,

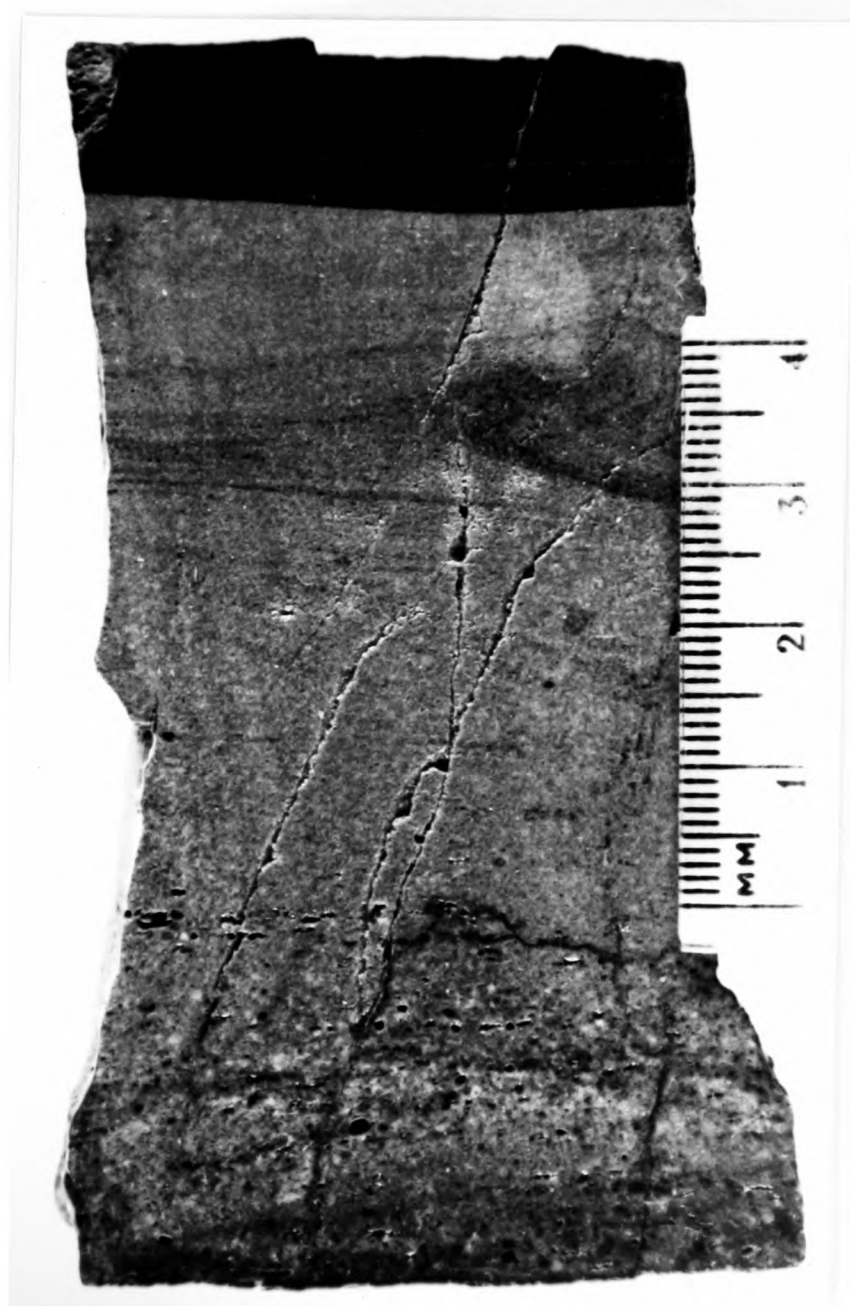


Fig. 2.4Cc

Pale bed from calcareous siltstones with coarse material in coarse basal parallel laminae, central part of bed rippled and convoluted. Sharp change from rippled sand to dark siltstone at the top. Unit 8 . Loc.353.



Fig. 2.4Cd

Calcareous siltstone bed with ripple lamination showing frequent truncation of the ripple laminae. Ripple amplitude increases from the base of the bed to the centre, and then decreases until the weak parallel lamination and dark graded siltstone at the top of the bed is reached. Unit 8. Loc.357.

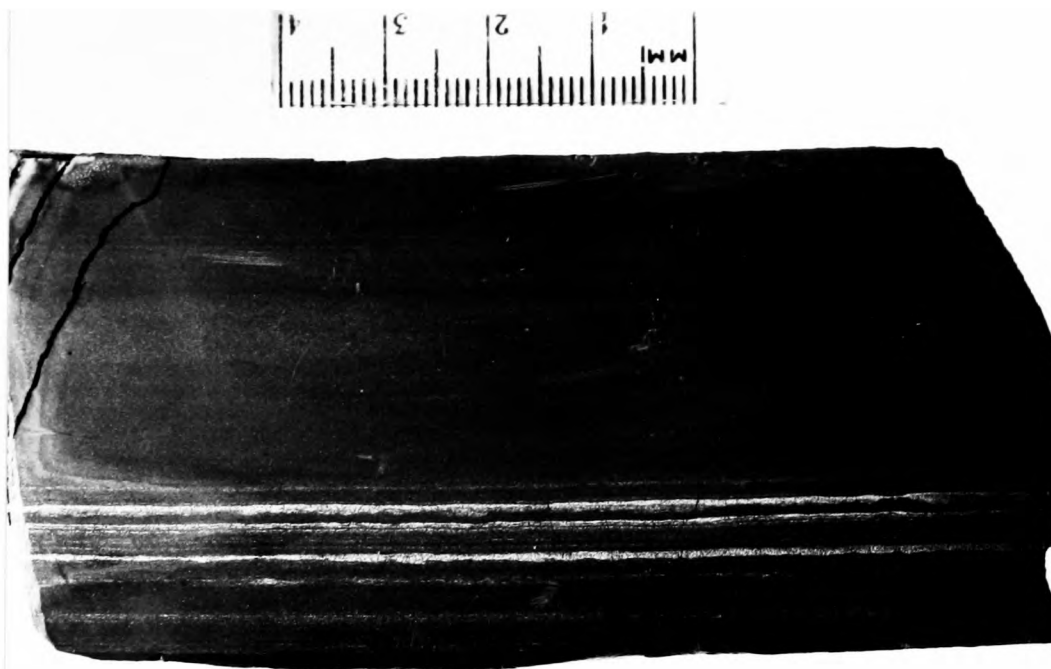


Fig. 2.4Ce

Typical dark calcareous siltstone bed with fine parallel quartz-rich laminae in the lower part of the bed. Unit 2. Loc.105.

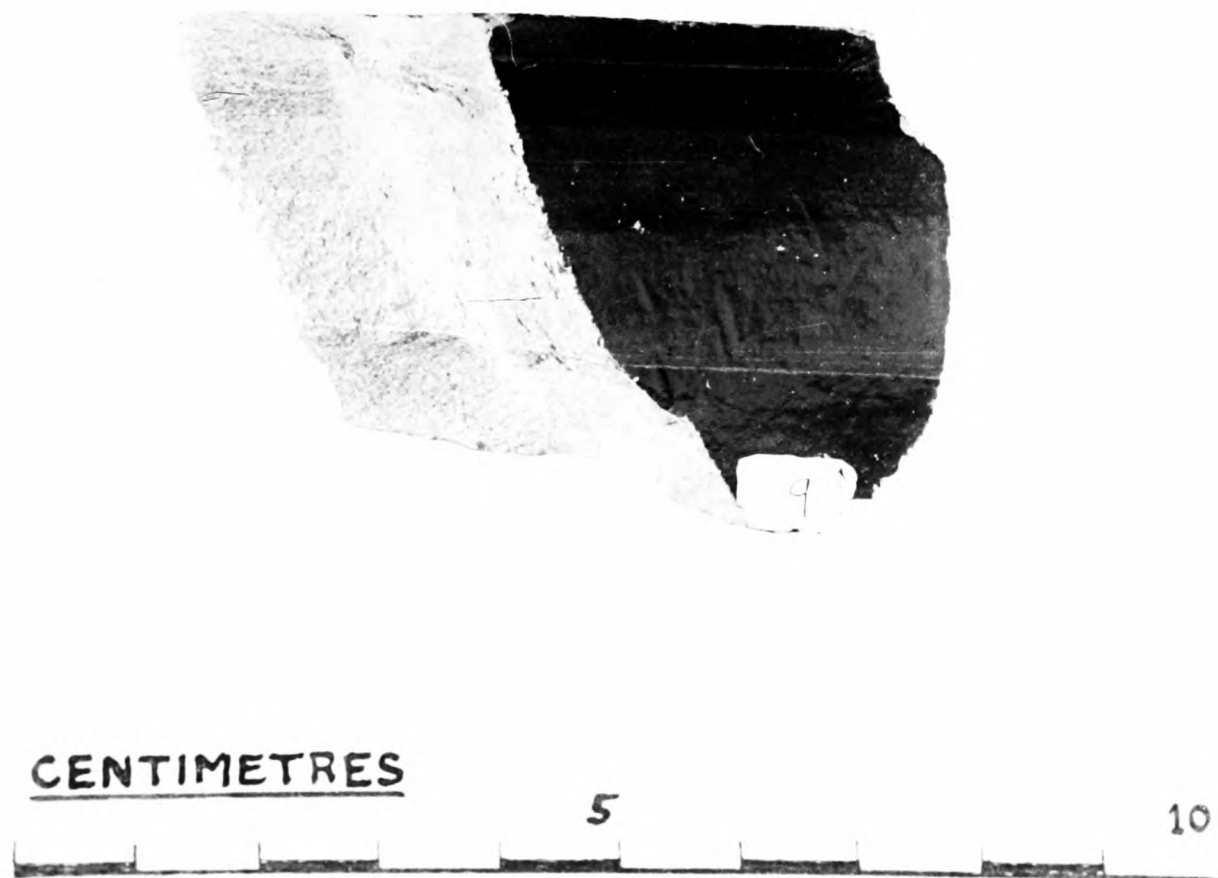


Fig. 2.4Cf

Example of the weakest of the calcareous siltstone beds with a single thin quartz-rich lamina at the base of the bed. Unit 0. Loc.055.

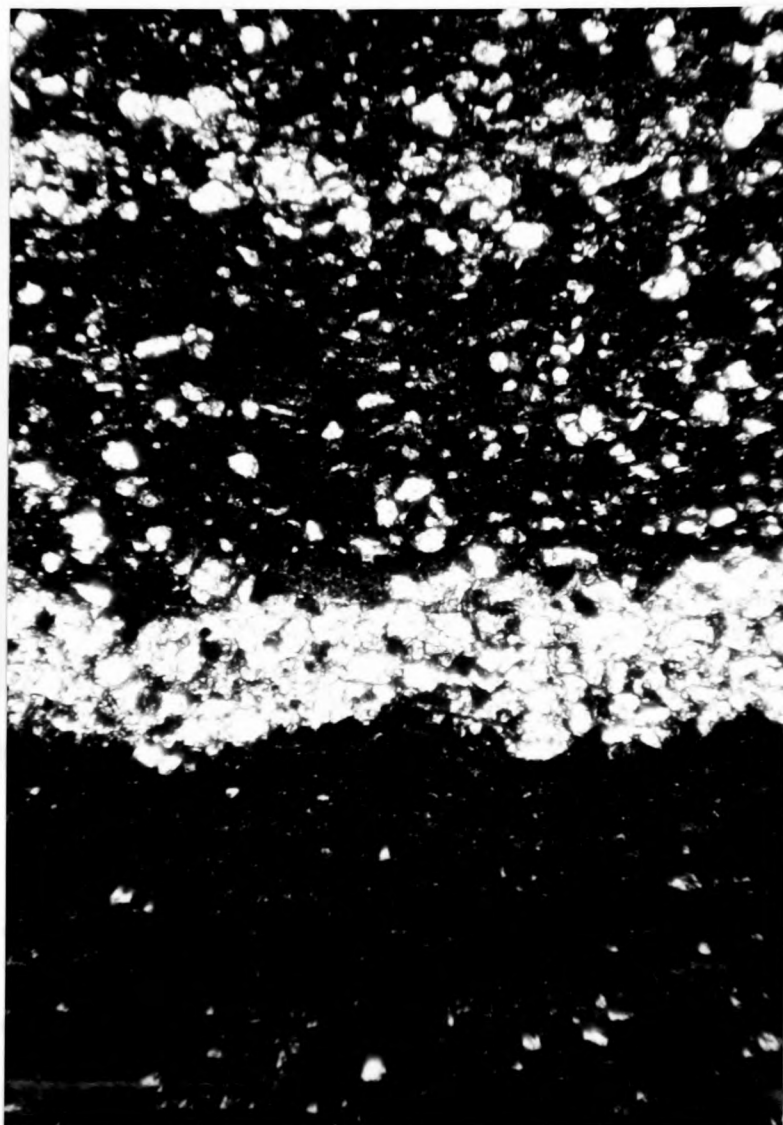


Fig. 2.4Cg

Basal quartz rich lamina of calcareous siltstone bed followed by dark poorly sorted siltstone. Bed rests on carbonaceous shale. x35, Unit 6 . Loc.055.

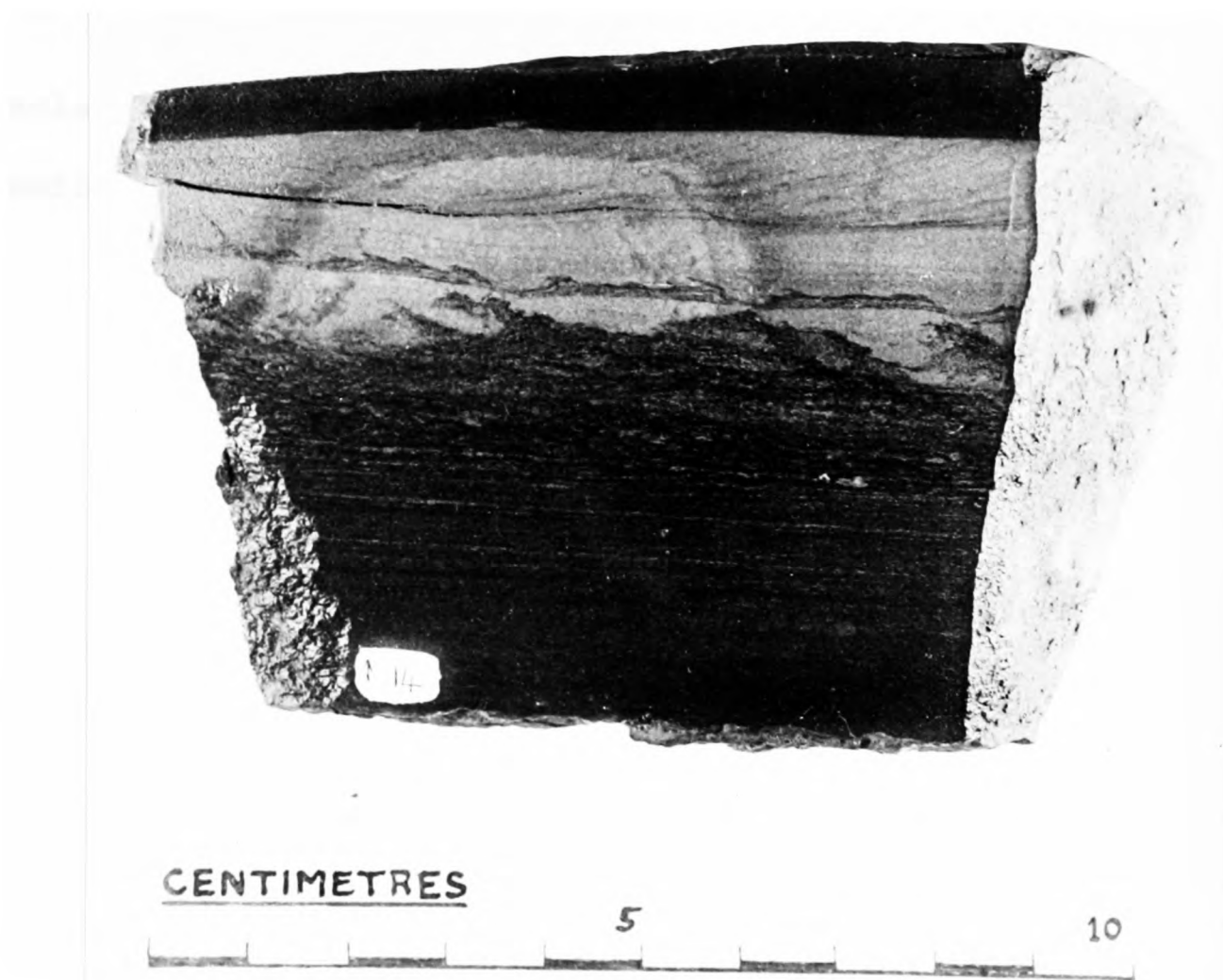


Fig. 2.4Ch

Dark calcareous siltstone bed with basal quartz rich lamina.
Top part of bed consists of ripples of quartzose sand with
a sharp top against dark siltstone. Unit 8 . Loc.008.

Fig. 2.4Ci

Isolated loaded ripples in dark siltstone and a quartz rich lamina with sharp top and disturbed base. Unit β . Loc.153.

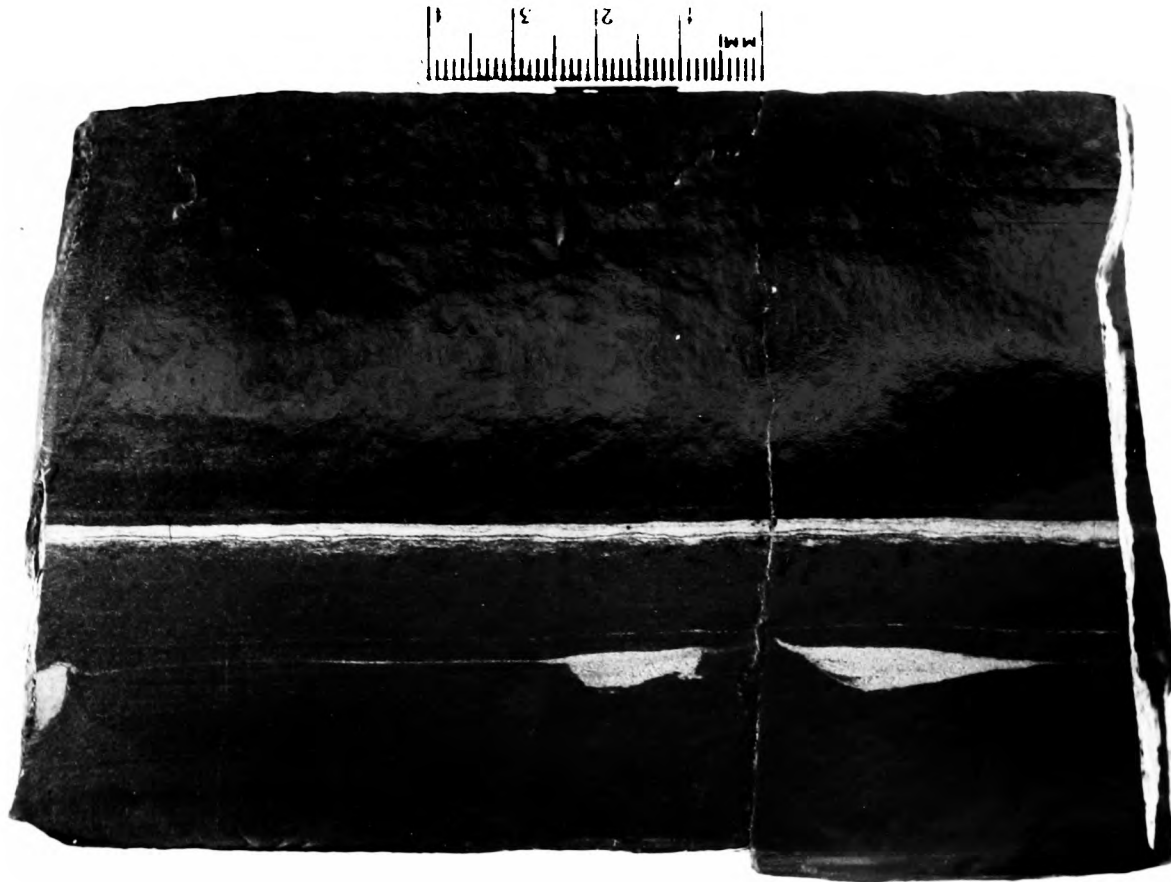




Fig. 2.4Cj

Isolated sandy ripple set in a dark siltstone bed with a pale basal lamina. Unit β . Loc.102.

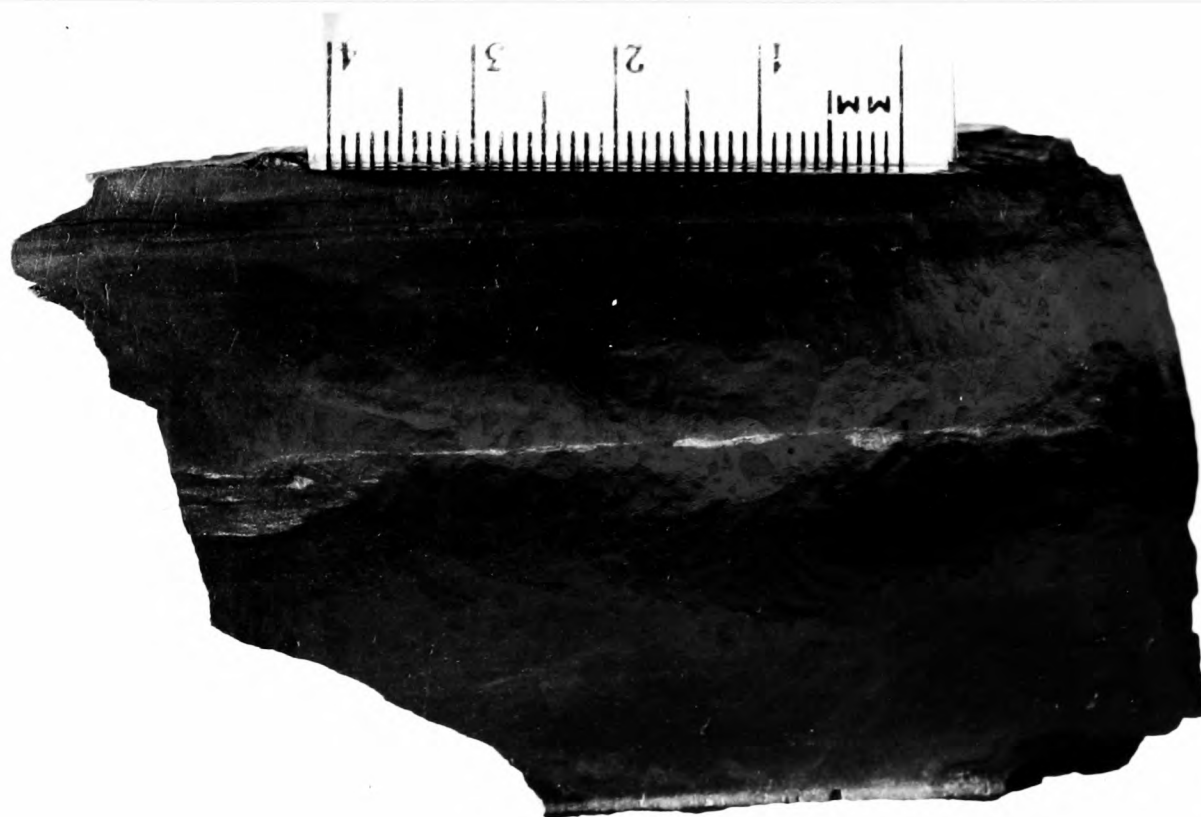


Fig. 2.4Ck

Bed of dark siltstone with a basal quartz rich lamina and a highly distorted quartz rich lamina within the bed.
Unit γ . Loc.353.

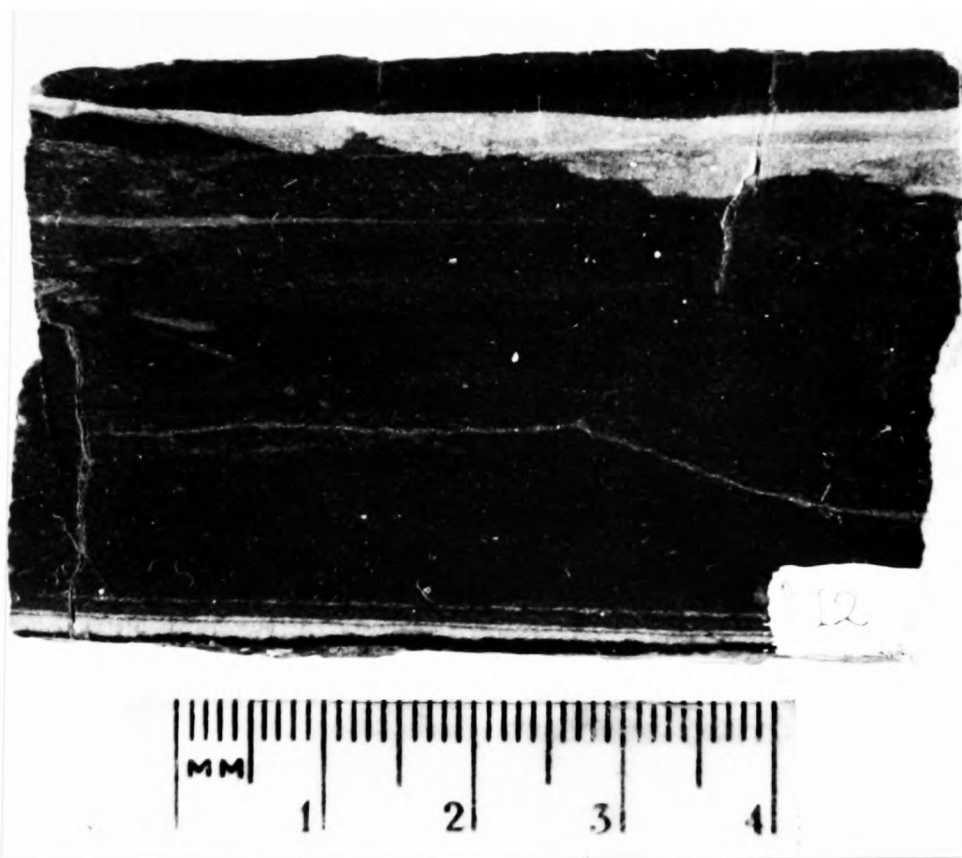


Fig. 2.4C1

Dark calcareous siltstone bed with fine quartz rich laminae at base, and loaded ripple of quartzose sand at the top. Unit γ . Loc.021.



Fig. 2.4Cm

Calcareous siltstone bed with ripples at base passing up into fine parallel laminae and dark graded siltstone. Unit unknown, near Loc.120.



Fig. 2.4Cn

Calcareous siltstone bed with rippled base followed by fine parallel lamination and dark graded siltstone. Unit ³. Loc.153.

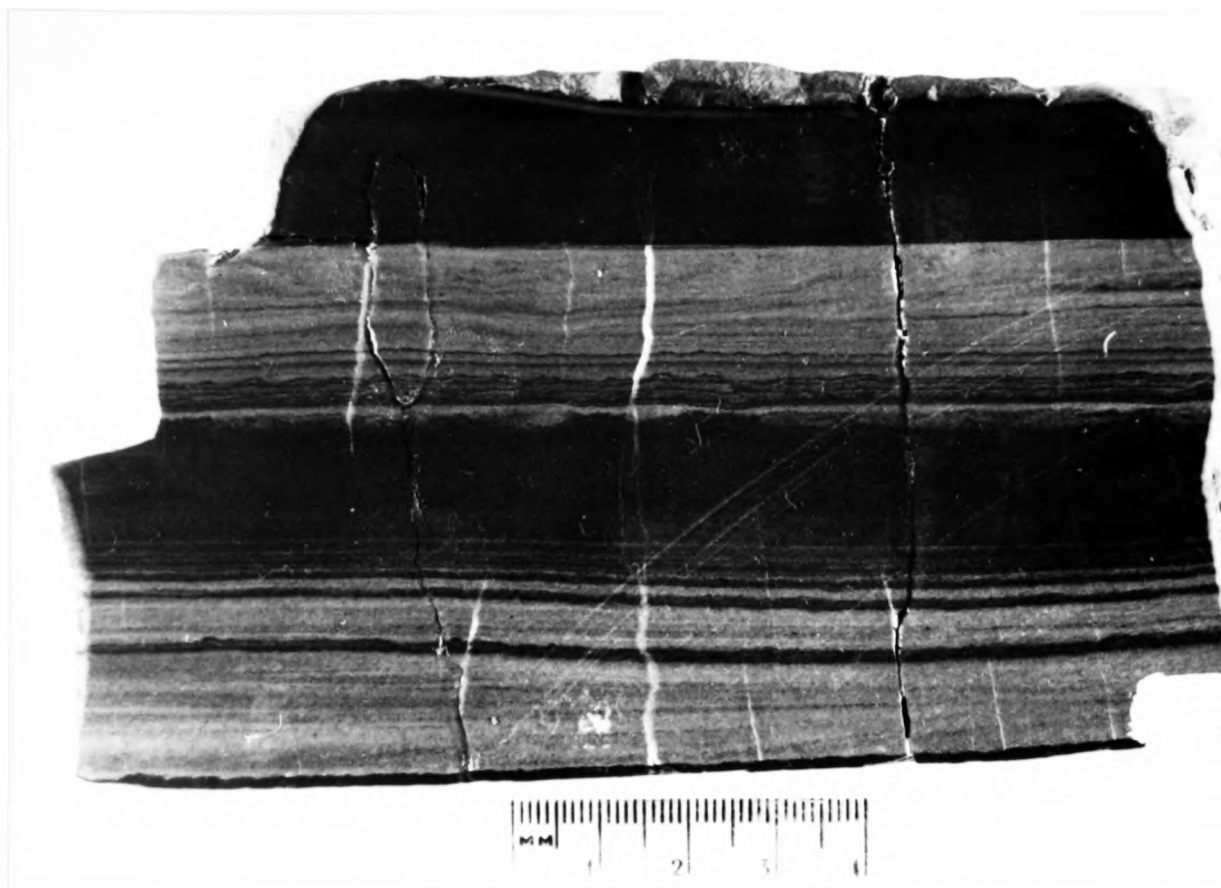


Fig. 2.4Co

Base of bed with very low angle ripple lamination followed by parallel lamination. Distortions in pale quartz silt near top of bed. Unit unknown. Near Loc.120.



Fig. 2.4Cp

Calcareous siltstone bed showing sequence of coarse parallel laminae — ripple laminae — fine parallel laminae. A row of small sunken sand ripples in dark siltstone at the top. Unit . Edale, Derbyshire.



Fig. 2.4Cg

Calcareous siltstone bed with coarse basal lamina followed by dark siltstone containing sharp topped pale laminae. Quartz rich laminae associated with flame structures near top below dark graded siltstone. Unit ². Loc.153.

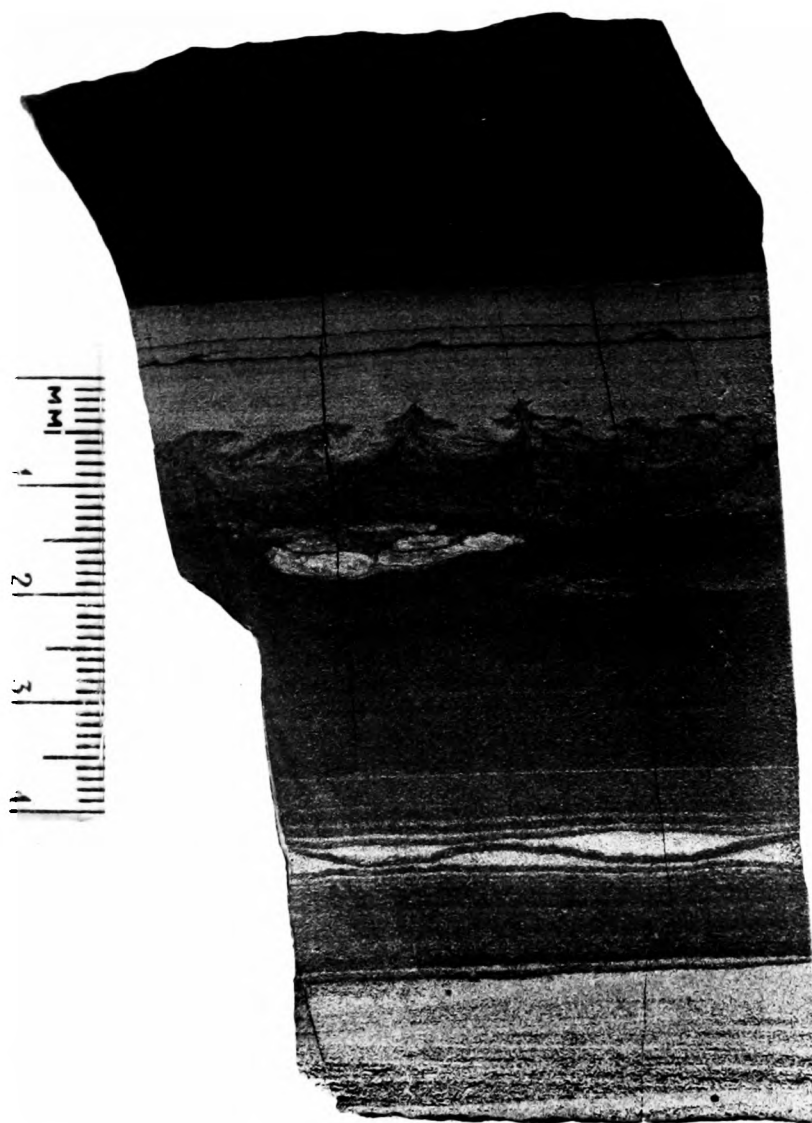
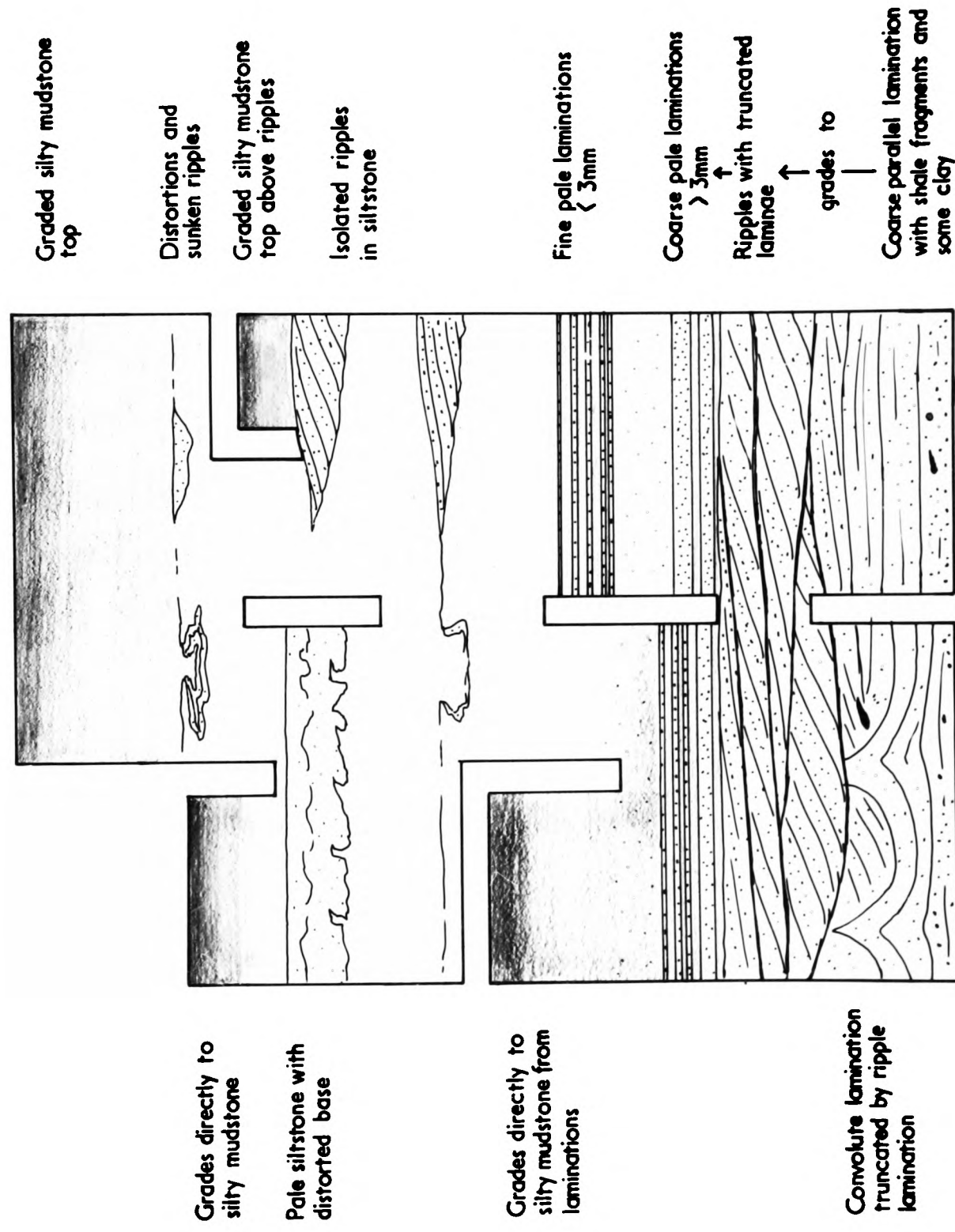


Fig. 2.4Cr

Calcareous siltstone bed with quartz rich base with very low angle ripple laminae. Small ripples included in lower part of dark siltstone and typical pale quartz rich laminae near top with flame structures at their base. Unit 3. Loc.153.

Fig 2.4Cs

VERTICAL SEQUENCE OF STRUCTURES OBSERVED IN THE
CALCAREOUS SILTSTONE BEDS



2.4D

FAUNA ASSOCIATED WITH THE CALCAREOUS SILTSTONES.

Evidence of a marine fauna is seen both within the calcareous siltstone beds and within the shales which separate the beds. The two situations will be considered separately.

Fauna between siltstone beds.

This fauna will be discussed further in the section on marine band sediments but it is relevant to consider the main features at this stage. The shale between the calcareous siltstones frequently contains the lamellibranch Posidonia corrugata. Specimens are usually represented by complete single valves which are usually seen on the bases of the calcareous siltstone beds. One single case was observed at Loc.109 Pyeclogh where the elongate valves were aligned on the base of a bed indicating some current orientation, but usually no orientation was apparent. Posidonia was generally found to be most abundant near the bases of Units β and δ which are underlain by goniatite bearing marine bands. No other species have been seen in the shales between the calcareous siltstones with the exception of rare crushed thin shelled goniatites (Anthracoceras or Dimorphoceras).

Fauna within siltstone beds.

Within the siltstone beds organic debris is

frequently seen and can be divided into two groups. The first group consists of material probably derived locally from the substrate over which the currents depositing the siltstone flowed. Whole valves of P. corrugata were probably derived in this way.

The second group consists of fragments of organisms which have not been found within the shales and probably have been carried a considerable distance to the site of deposition. The most noticeable elements in this group are crinoid ossicles and fragments of fenestrate bryozoans. The fragments seen are of a comparable grain size to the sediment containing them, and were undoubtedly acting as detrital grains. Endothyrid foraminifera are of frequent occurrence and with them are often seen fragments of other organisms which cannot be identified. The foraminifera are frequently pyritised (Fig. 2.4 Ig), (Holdsworth thesis 1963 Pl. 25(1), 26, 27).

Present within the siltstone beds are siliceous spicular structures often filled and replaced by pyrite, which are possibly sponge spicules (Ibid p.210). Bullions (calcareous concretions) from marine horizons frequently contain both radiolaria and sponge spicules, and thus these elements of the fauna were living within the basin area and do not necessarily form part of the derived fauna. There is no evidence that the crinoids, bryozoans or

endothyrid foraminifera were living in the basin area at the time of deposition of the siltstones and their absence from bullion limestones (Holdsworth 1966 p.321) confirms this view. This feature combined with the fact that the organic fragments are of comparable grain size to that of the sediment which contains them points to a derived origin. Thus the source area of the sediment comprising the calcareous siltstones was probably a marine area of shallow water. As the detrital minerals appear to be derived from a land area similar to that supplying the protoquartzitic material of the turbidites (see section 2.4I), the calcareous siltstones could be derived from a shallow marine area, possibly a shelf area, bordering such a land mass.

2.4E

CURRENT DIRECTIONS.

Very few current directions could be determined for the calcareous siltstones since the normal features from which current directions are determined are absent or not favourably exposed. Bottom structures are absent, and the ripples that are present are often not detectable in outcrop and could only be seen when the specimens were cut and varnished. The ripples have never been observed exposed in plan view, but they

are certainly not straight crested and directions determined from them would give only general current directions.

Flame structures at the bases of isolated ripples have been observed which are presumably overturned in a down current direction (Fig. 2.4Ch) and in several cases a general direction has been determined from these structures. In exposures of Unit γ at Sunnydale, Elkstones and Croker Hill flame structures indicate transport from the south or south-east. These determinations are few in number and little significance can be attached to them. The determinations were made on isolated rows of ripples which may represent reworking currents and not the direction of initial supply of material to the basin area.

Attempts at determining grain orientation in the basal laminae of the siltstones were not successful due to the secondary and replacive carbonate present and the presence of secondary silica. Both these factors made determination of original grain shape, and hence orientation, impossible.

2.4F

THICKNESS VARIATIONS OF CALCAREOUS SILTSTONE UNITS.

The variation in measured thickness of Units β and γ are given in Fig. 2.4Fa. Unit α has not been

studied in detail but is known to be thick and apparently uniform throughout most of its thickness. Unit α corresponds with the 'Crowstones' of Hudson (in Hudson and Cotton 1945a p.321) which were considered to be 550 feet (c160m.) thick in the Gun Hill Borehole.

Nowhere in the field can units of this thickness be measured with any confidence and this is the most reliable figure known for Unit α in the basin area. At the base of the unit in the Gun Hill Borehole Cravenoceras leion Bisat (E1a) was tentatively identified (Hudson in Hudson and Cotton 1945a p.321). Within the basin area an exposure with C. cf. leion occurs at Loc 260 at the presumed base of Unit α just above the Onecote Sandstone, but no indication of the absolute thickness of Unit α can be gained from this area.

The variations in thickness of Units β and γ shown in Fig. 2.4Fa are surprisingly small considering the distances apart of the localities. No regular variation in thickness can be detected and it can only be concluded that these units form a very constant cover over most of the basin area. Variations in thicknesses of individual beds within units at different localities have been observed but this has no marked effect on the absolute thicknesses of the units. This feature is in marked contrast to the protoquartzitic turbidites where a regular thickening of both individual beds and units could be

detected towards the source area.

The very even distribution of the calcareous siltstone over most of the basin area implies a very flat and uniform depositional area over which the available sediment was very evenly distributed.

In the Upper Dove the lowest horizon seen is the E2a2 faunal band and the E2b1 band can be observed only a few metres from the exposed plane of unconformity. Clearly the calcareous siltstone units do not transgress onto the limestone massif area, and have not been observed resting against the unconformity surface although they may do so at depth. The absence of the calcareous siltstones at the basin margin could be due to thinning - as observed in the protoquartzites -, overlap of the Namurian onto the massif, or a combination of both mechanisms.

Fig. 2.4Fa.

<u>Locality Area.</u>	<u>Unit β Thickness cms.</u>	<u>Unit γ Thickness cms.</u>
Hurdlow	1400	520
Croker	-	800
Blake	1400	650
Pyecclough	1400	900
Elkstones	1300	650
Shirkley / Sprink	1300+	600
Winkhill	1000max	600
Edale	550	800

2.4G

AREAL DISTRIBUTION OF CALCAREOUS SILTSTONES.

The areal distribution of exposures does not give such good cover of the basin area as was possible in the case of protoquartzitic turbidites. The siltstones are not well exposed in the area to the west and north of the Leek where most of the available exposure in E2a is of the proximal protoquartzitic turbidites.

The majority of the localities where the siltstones are exposed display very similar developments which could not be subdivided. These areas are indicated on Fig. 2.4Ga* as having 'normal' calcareous siltstone developments which are typified by the log (Fig. 2.4Gb*) of the calcareous siltstones, and Figs. 2.4Ca-r of sectioned beds.

Variation between Units α , β and γ is difficult to detect, but in the field ripple lamination is apparently more frequent in Unit γ than in the lower units. This is particularly so in the case of rows of isolated ripples which are rare in Units α and β . No difference could be detected between Units α and β . Unit α has not been examined in detail, but it is usually poorly exposed. In contrast to the variation observed in the protoquartzites the calcareous siltstones are remarkably constant in their developments from Croker Hill to the north-west through Hurdlow, Thorncliff and Pyecloough to Elkstones in the

* Figure in folder.

south-east of the area. Only three successions exposed in the basin area show significantly different developments of calcareous siltstones worthy of comment.

In the extreme south-east of the area at Winkhill (Locs.581-6) small portions of all three units are exposed, and although the exposure is poor there is a marked reduction in strength of the individual siltstone beds which in Unit Y do not exceed 9cm. in thickness - in contrast with the normal 20cm. maximum. There is, however, no absolute reduction in the thickness of the unit relative to the 'normal' areas (Fig.2.4Fa), and the reduction in thickness of the beds reflects a greater proportion of fine grained material and reduction in abundance of ripple and parallel lamination. The absolute number of hard siltstone beds in Unit Y is about 25 at Loc.581 whereas in the 'normal' areas, e.g. Hurdlow, at least 60 hard siltstone beds are present in this unit. This feature could in part be secondary since it was frequently observed that weathering of thin dark siltstones reduced them to a shale texture especially when only poorly calcified.

The section through the Y siltstones at Uppper Shirkley shows a weaker development than at Winkhill. Only 14 dark siltstone beds are present in the unit and these have a maximum thickness of 6cm. Weak parallel

lamination is present and only a single row of ripples was seen. The proportion of ordinary shale in this section is 80% as compared with 20-50% in 'normal' developments of the calcareous siltstone. Here again the absolute thickness of the unit is roughly the same as that seen in the 'normal' areas, the difference being only in the strength and frequency of the hard siltstone beds. Calcification is generally weak at Shirkley being usually confined to the siltstone beds. Siderite is absent and the pale laminae contain the diagenetic pyrite typical of the calcareous siltstones.

The third abnormal development of the siltstones concerns Unit β at Sprink. Here the unit consists mainly of calcareous shale with abundant, but small, Posidonia corrugata. Only weak parallel lamination is developed and this is usually seen as a pale lamina at the base of a thin bed. The development is thus unusual in containing abundant small P. corrugata and having only weakly developed lamination.

Outside the north Staffordshire basin area developments of the calcareous siltstones have been seen in Edale (Part 3.2) and rocks considered to be probably equivalent to them occur at Astbury (Part 3.2). The Astbury section is considered to be on the western margin of the basin area and is a greatly thinned succession. It possibly represents the edge of a shallow water area

where the derived benthonic fauna of the calcareous siltstones was living, but there is no evidence of the fauna actually living in the Astbury area.

The Edale exposures show a fairly 'normal' development of the calcareous siltstones both in the thickness of the units and the development of sedimentary structures. It appears that the siltstones of Edale were derived from a similar, if not the same, area to that supplying the north Staffordshire siltstones. Such an area could have been to the west of the Staffordshire and north Derbyshire basins on a 'shelf' area of which the Astbury section may represent the edge. This is the most likely source area of the detritus in the calcareous siltstones. The material described by Kent (1948) from a borehole at Formby, Lancs. may possibly represent the source material of the calcareous siltstones (see Part 4). It is unlikely that the material was derived from the south since this is the direction in which the siltstone beds have been observed to reduce in strength and lack coarse parallel laminae and the ripple laminae. Similarly, if the Staffordshire siltstones were derived from the south a separate source would have to be found for the Edale siltstones. Derivation from the east is most unlikely since the Derbyshire massif area lay in this direction and on it in E2a is found a very thin reduced succession

without significant detrital material of coarse silt or fine sand grade (Ramsbottom et al. 1962). Derivation from the north is unlikely for the Staffordshire siltstones as they would have to be transported around the end of the massif area, and developments to the north of the massif would be expected to be stronger than those in the south away from the source area.

The few tentative current directions presented do not indicate currents from the west but rather from the south. These current directions are based on rows of isolated ripples which have been interpreted as due to reworking, and thus they may not indicate the direction of the source area, but only the direction of a reworking current.

2.4H

MINOR DEPOSITIONAL CYCLES WITHIN THE CALCAREOUS SILTSTONES.

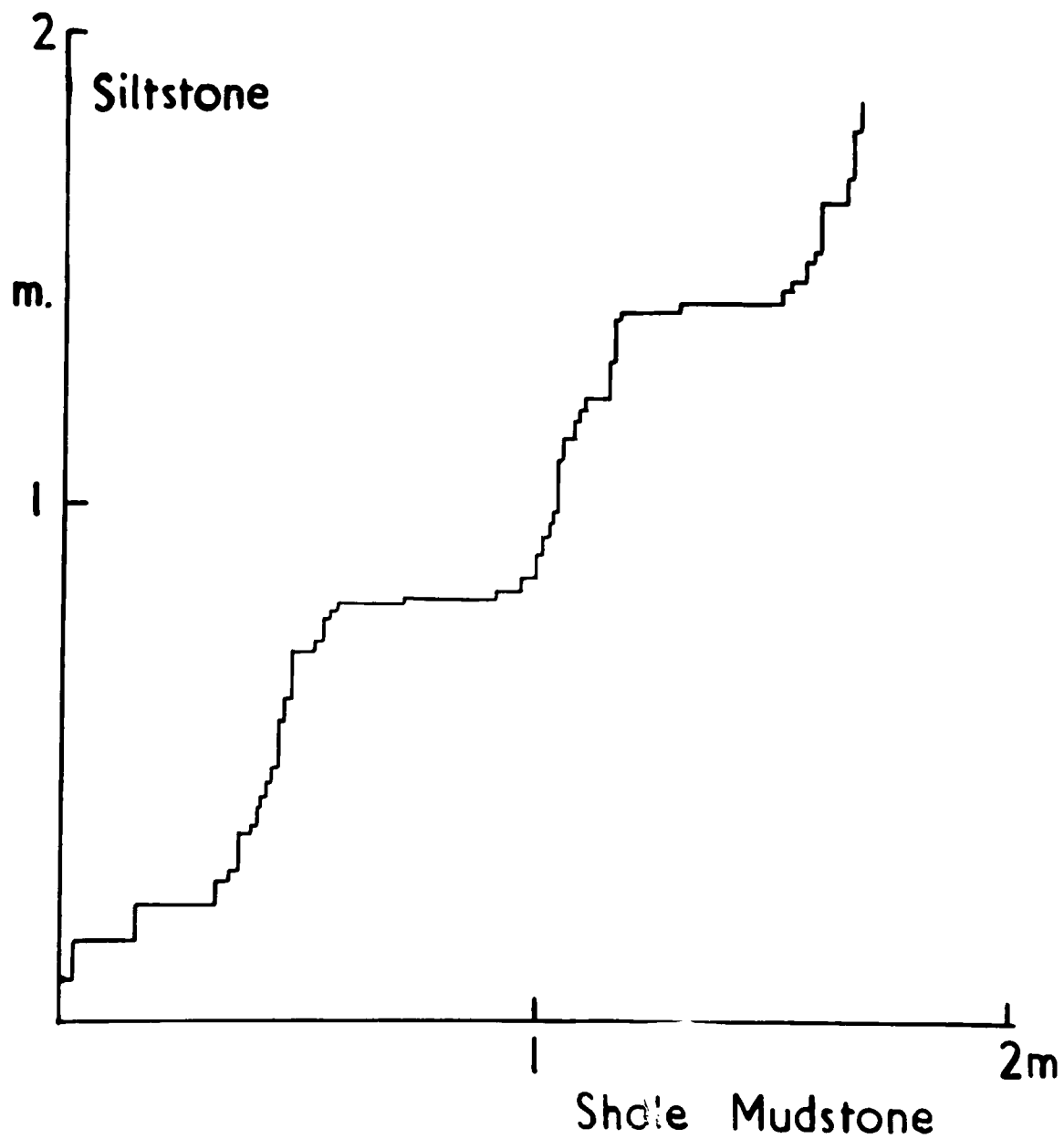
It has already been noted that within the proto-quartzite lithofacies the thicker turbidite beds have a tendency to group together. Within the units of calcareous siltstones the same feature is observed but is not so obvious from outcrops probably due to the weathering of the siltstones.

The log (Fig. 2.4Gb.) of the calcareous siltstones Unit β of Oakenclough Loc 153 shows the grouping of thick beds and a cumulative shale/siltstone diagram (Fig. 2.4Ha) illustrates the minor cyclicity present in this lithofacies. Not all sections of the calcareous siltstones examined show such a regular cyclicity and when siltstone/shale diagrams are plotted no regular cyclicity is apparent. This is particularly so in the case of Unit γ , the thinnest of the calcareous siltstone units. No long fully exposed sections are available in the calcareous siltstones to examine this apparent minor cyclicity over a reasonable thickness of strata.

FIG 2.4Ha

Cumulative siltstone/shale-mudstone diagram.

Minor cycles in Unit β Calcareous Siltstones.
Oakencloough Loc. 153.



2.4I

PETROGRAPHY OF THE CALCAREOUS SILTSTONES.

There is considerable variation within the calcareous siltstone beds particularly in respect of the proportion of dark 'matrix' material present. The pale laminae and ripples generally lack dark matrix material and consequently are better sorted than the dark siltstone parts of the beds.

The overall mineralogy will be considered first and then the relative composition of the dark and light laminae.

DETRITAL MATERIAL.

Quartz.

Quartz is the major detrital mineral present, individual grains in the coarser parts of a bed have been observed up to 1mm. in diameter but usually the maximum grain size is in the range 0.5mm. to 0.15mm. at the base of a thick bed. The grains are identical in type to those seen in the protoquartzites as far as can be judged, and are probably derived from an area of similar rocks to those which supplied the material for the protoquartzites. The quartz rich fragments of a brownish colour described from the protoquartzites section 2.1G, and thought to be of volcanic origin are also present in the calcareous siltstones.

Feldspar,

K-feldspar is absent from the calcareous siltstones, and only trace quantities of plagioclase are present. The plagioclase is present as fresh grains close to albite in composition. The feldspar content is thus identical with that of the protoquartzites.

Muscovite Mica.

White mica is usually concentrated in the upper parts of the dark siltstone beds where it may constitute over 5% of the rock (Fig. 2.4Im). Usually it is present only as isolated highly birefringent plates of detrital origin, which are sometimes observed to be in the process of alteration to kaolinite (seen in sections stained with methylene blue).

Heavy Minerals.

In thin section grains of tourmaline and zircon are seen which are of the same type as those seen in the protoquartzites.

Clay Minerals.

Sections stained with methylene blue show kaolinite to be present in some beds sometimes constituting nearly 10% of the rock, and usually occurring as well crystallized patches interstitial to the quartz grains. Undoubtedly much of the kaolinite is secondary but it is possible that some detrital kaolinite is present.

Holdsworth (thesis 1963 p.191) found no evidence of well crystallized kaolinite in stained thin sections. It is possible that the kaolinite has a variable distribution over the basin area but insufficient evidence is available at present to trace such a variation. No other clay minerals can be determined by microscopy but diffractometer traces of dark siltstones reveal the presence of small amounts of mixed layer clay material of mica-montmorillonite type. The material is very irregularly interstratified.

Shale Fragments.

Shale fragments are seen included in the lower parts of the thicker beds (Figs. 2.4Cb, 2.4Ia). They are similar in composition to the underlying shale substrate. Frequently they can be observed to be in the process of disintegration to fine grained matrix material and it is possible that a large proportion of matrix is derived from the breakdown of shale fragments of various sizes.

Organic Debris.

The derived fauna of shelly debris including crinoids, bryozoans and foraminifera has already been described (section 2.4D), but of far greater abundance in the siltstones is dark opaque or sub-opaque organic material. It is not possible to determine the source of much of the material which now appears to be very finely divided

carbonaceous material. Some elongate carbonaceous flakes may represent carbonised plant fragments, but macroscopic plant material is absent from the calcareous siltstones.

SECONDARY MATERIAL.

Carbonates.

Holdsworth (thesis 1963 p.192-200) investigated the carbonates present within the calcareous siltstones and concluded that calcite, ankerite and dolomite were present in variable proportions, and that siderite was absent or only present in trace quantities. Testing of cut faces with dilute HCl shows that there is considerable variation in the proportions of the carbonates present. Some beds are rich in calcite and others contain none.

In thin section the different carbonate phases cannot usually be identified but one section was observed with well crystallized rhombohedral crystals which gave no reaction with dilute HCl and appeared to represent carbonate of dolomite-type. The carbonate is all secondary (with the exception of calcareous fossil fragments) and is usually seen to be replacing detrital quartz, (Fig. 2.4Ib) or both quartz and matrix material. Fine grained carbonate in the matrix is difficult to distinguish and thus the proportions of carbonate counted in thin section possibly are low. There is a wide variation in the total carbonate present as shown in Fig. 2.4Im. The maximum amount may

exceed 50% of the rock in pale laminae which are extensively replaced. The proportion of carbonate present does not appear to affect the overall colour of the rock.

Quartz and Chert.

Considerable redistribution of silica has taken place in the calcareous siltstones and secondary silica is present in two forms. Clear quartz rims to original detrital grains are frequently developed (Fig. 2.4Ic) and give straight contacts between detrital grains. More important than optically continuous quartz rims to grains is the development of chert within the calcareous siltstones. In coarser beds it may be evident as cherty cement to quartz grains (Fig. 2.4Id), but in the matrix the development of fine secondary chert is probably of greater importance, though the sub opaque nature of much of the groundmass makes recognition of secondary chert difficult.

Originally calcareous fossil fragments have been observed preserved as chert (Fig. 2.4Ie), but it is not possible to be certain that they were not silicified before deposition. The obvious presence of secondary silica in the rocks makes it likely that the replacement of organic calcite by silica occurred after deposition.

Pyrite.

Pyrite of a secondary nature is frequently seen in

the calcareous siltstones, and it occurs in three different forms.

Some pale bands contain patches of pyrite clearly visible in hand specimen and frequently the pyrite is concentrated towards the middle of a pale band. The pyrite is well crystallized and clots of crystals probably of cuboid habit are often seen (Fig. 2.4If). The pyrite may locally constitute 15% of a pale lamina.

Pyrite is also present in a framboidal form particularly within the dark siltstone or associated with microfossils (Holdsworth thesis 1963 p.210 Pl. 25(1), 28, 29). Actual fossil fragments, particularly foraminifera, are often replaced by pyrite (Fig. 2.4Ig). Some beds contain only rare pyritised fossils but in others such fragments are common (Fig. 2.4Ih). In such cases the foraminifera fragments are clearly favoured for pyrite replacement.

Kaolinite.

Patches of well crystallized kaolinite, similar to those seen in the protoquartzites, are of secondary origin and are most clearly seen in the pale and coarser basal portions of the beds. In the dark siltstone kaolinite is also present, but in a more finely divided state, and it is not possible to determine whether or not it is of a secondary origin.

MODAL COMPOSITION OF THE SILTSTONE BEDS.

Modal analyses are presented in Fig. 2.4Im of four typical parts of the calcareous siltstone beds. They illustrate the wide variation found within the individual parts of the beds, particularly with respect to matrix content.

The most important point to be stressed is that the modal composition bears little relation to the composition of the originally deposited sediment. Slide 353.12B is a good illustration of this point, since over 50% of the material now present is recognised as being of secondary origin. Despite this great deficiency a few points do deserve attention.

The lower laminae contain matrix material in the form of kaolinite, carbonaceous material and fine grained clays etc. Shale fragments are also frequently present and are similar in composition to the shales separating the calcareous siltstone beds. The quartz present ranges from coarse silt to fine or medium sand grade, (Wentworth (1922) in Pettijohn 1957 p.18) with rare grains up to 1mm. (coarse sand) present. The average grain size of the quartz in the sediment probably lay in the fine to very fine sand grades, but accurate analysis is not possible due to the presence of secondary silica.

The rippled laminae (Spec. 353. 12B) are better sorted than the lower laminae, lack dark matrix and have reduced amounts of kaolinite, shale fragments, and carbonaceous material. Carbonate is usually abundant and probably partly fills original voids, but also extensively replaces detrital quartz. The quartz grains are better sorted than in the lower laminae and are usually of fine to very fine sand grade. The lithology of both the lower laminae and the ripple laminae could be regarded as fine calcareous quartzose sandstone. Upwards within the beds the quartz grain size decreases and the upper pale laminae contain quartz predominately of silt grade (less than .625mm.) and are thus calcareous quartzose siltstones (Holdsworth thesis 1963 p.192). Typical matrix free material is shown in Fig. 2.4I. j.

The dark matrix-rich layers of the beds (typified by slides 353.12C and 353.11) contain from 20% to 40% of detrital quartz of silt grade set in the dark matrix material, along with mica flakes and carbonaceous fragments.. Within the matrix the grain size of the originally deposited material cannot be determined with certainty, but a considerable proportion of it is probably of silt grade. Certainly most of the micaceous flakes and many carbonaceous fragments fall in this category. Clay grade material is also present both in the form of clay and carbonaceous material and in the finer parts of the beds it may exceed

the silt grade material. It is considered that the majority of the dark layers are predominantly silty and the name calcareous siltstone or calcareous muddy siltstone (Fig. 2.4Ik) would be appropriate. The tops of the beds with clay in excess of silt are generally silty mudstones (Spec.353.11) and when calcified-as is often the case-calcareous silty mudstones (Fig. 2.4Il).

The bulk of the hard beds present in all developments of the lithofacies approximate to the composition of calcareous siltstone and thus this term is used for the lithofacies.

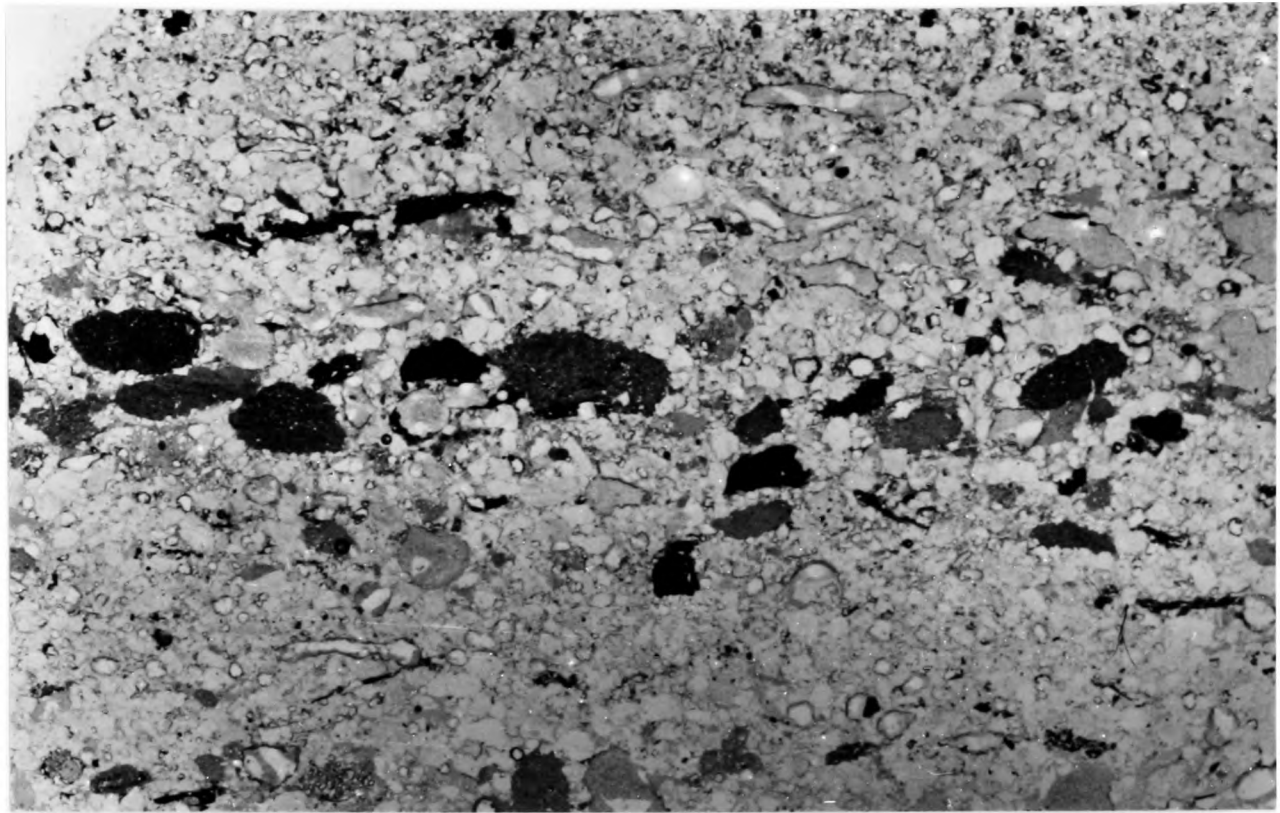


Fig. 2.4Ia

1 mm

Small rounded shale fragments included in the base of a calcareous siltstone bed. Slide 353.6. Unit 8. Loc. 353.

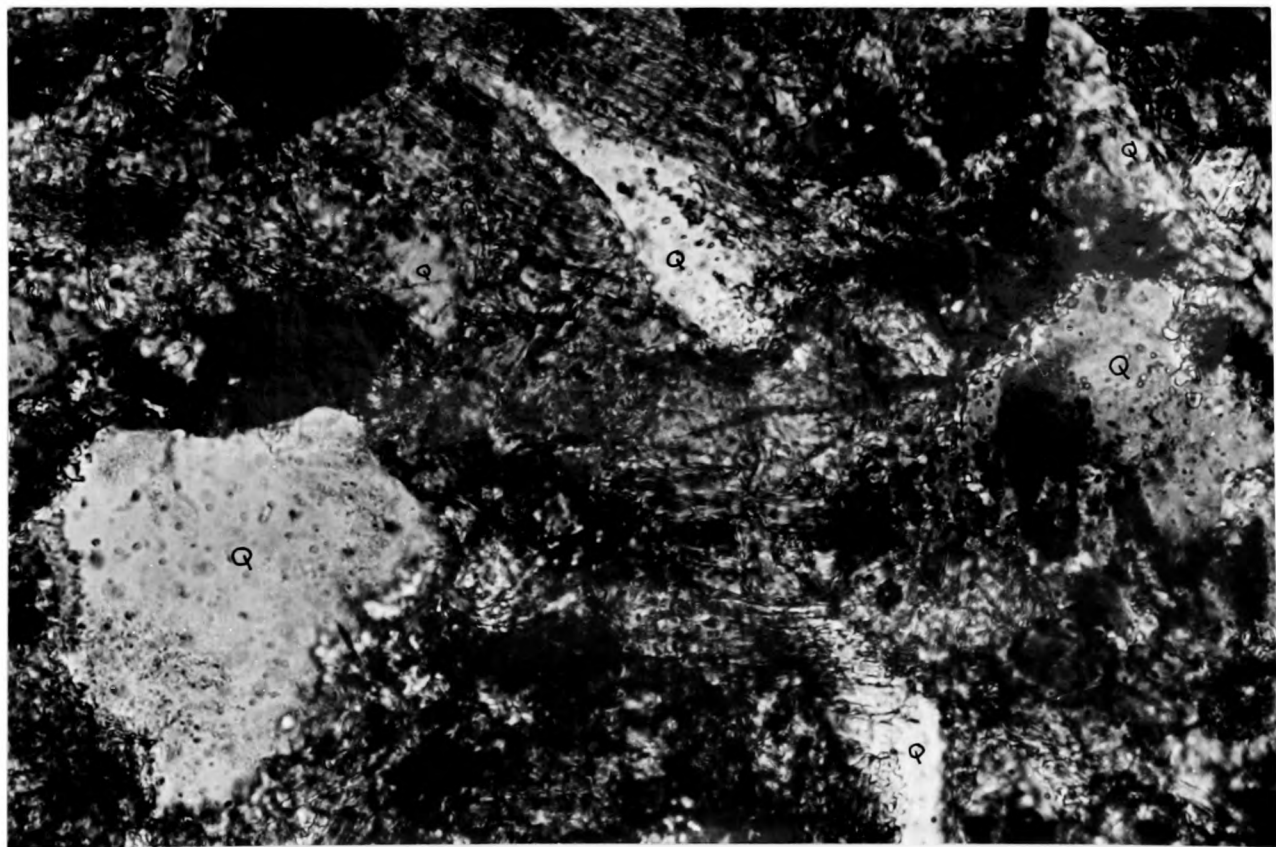
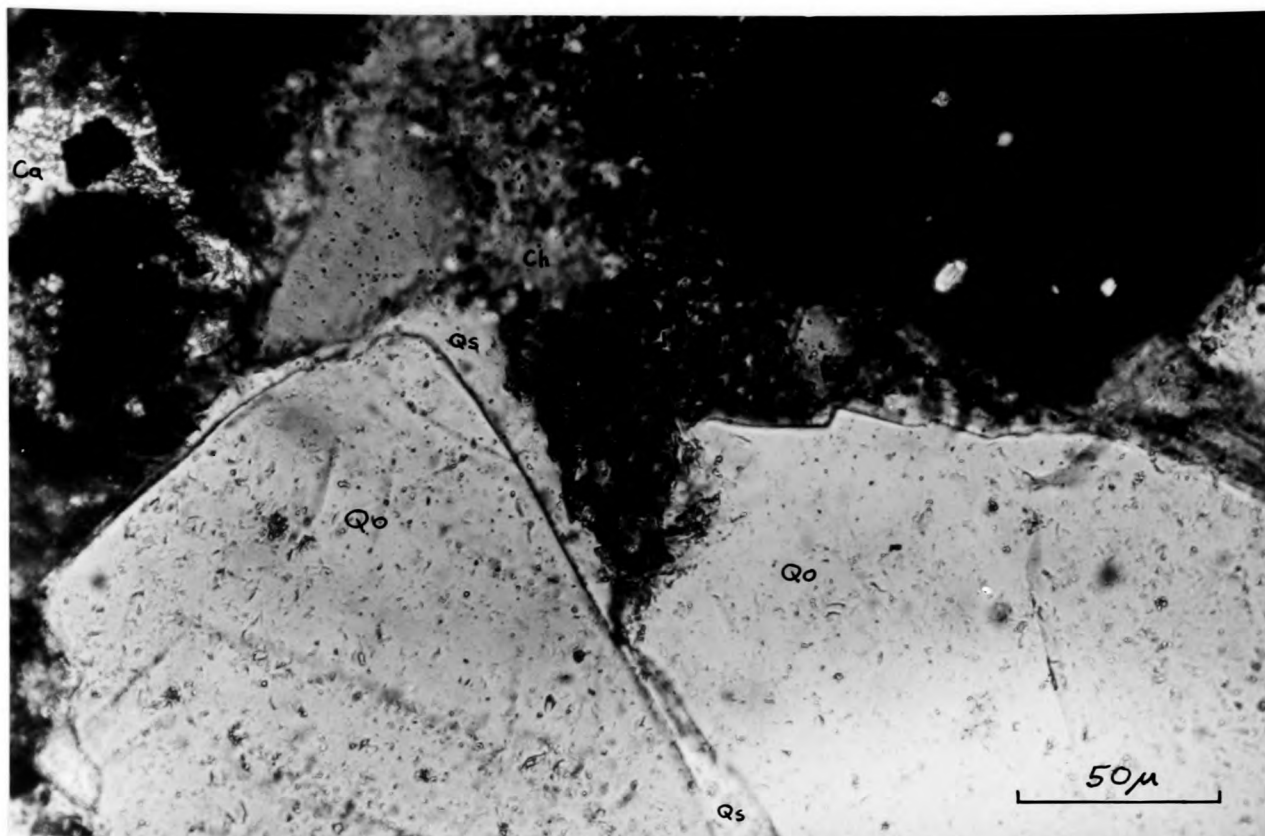


Fig. 2.4Ib

50 μ

Calcite replacing quartz (Q) in a pale siltstone bed. Slide 109.3b. Unit ? Loc. 109.

Fig. 2.41c



Secondary quartz rims (Qs) to original quartz grains (Qo) with Chert (Ch), Vermicular kaolinite (K) and Calcite (Ca) in a calcareous siltstone bed. Slide 205.1d, Crossed polars. Unit 8. Loc.205.

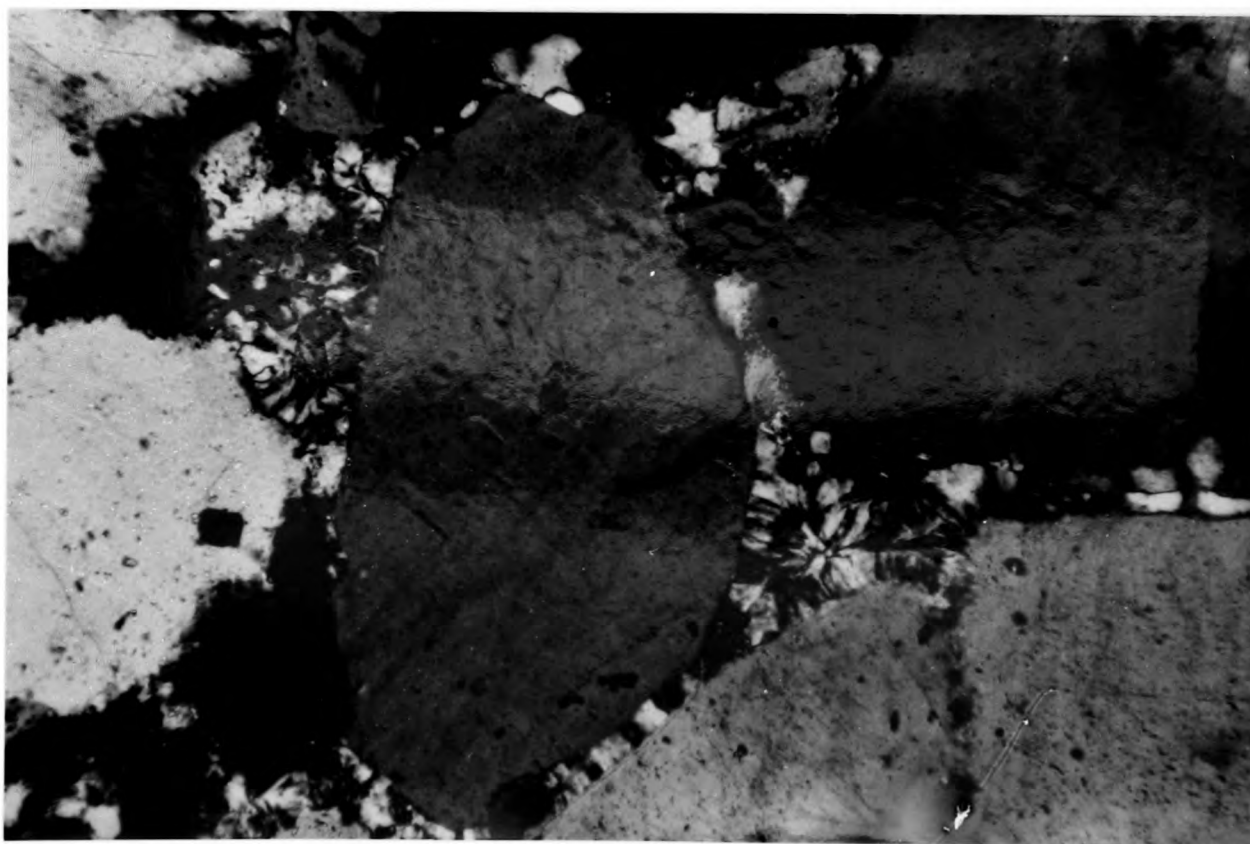


Fig. 2.41d

Chert cement to quartz sand at base of a calcareous siltstone bed. Slide 353.6a, Crossed polars. Unit 8. Loc.353.

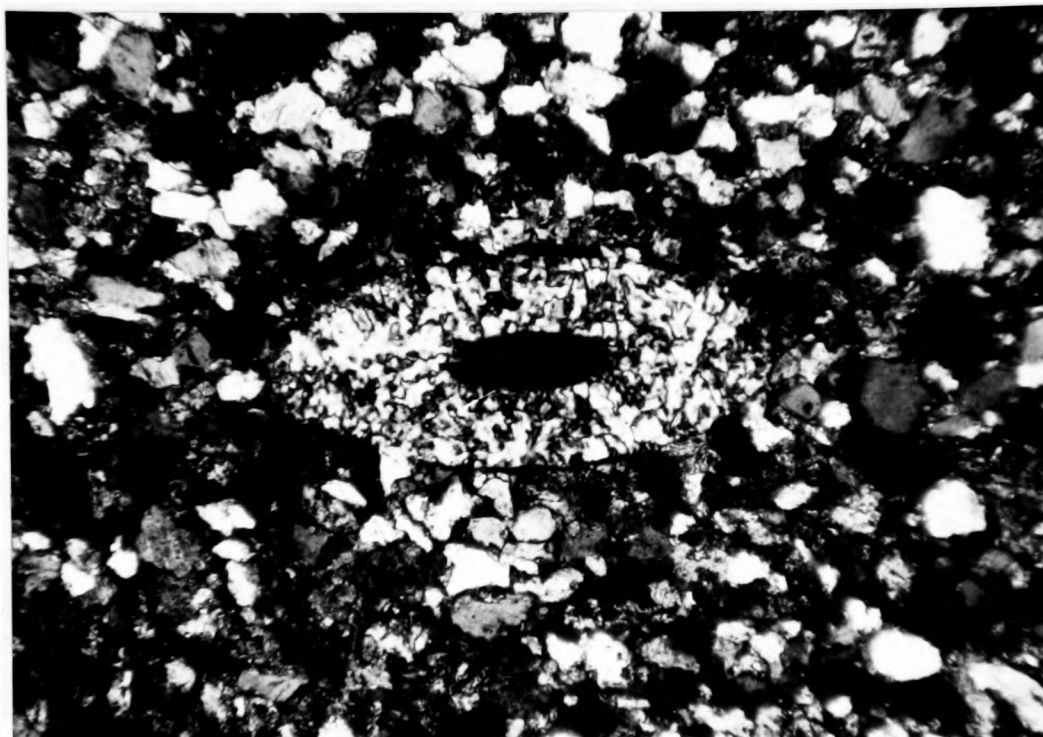


Fig. 2.4Ie

100 μ

Section of silicified spine, possibly from a brachiopod, in calcareous siltstone. Slide 353.12, Crossed polars. Unit 8. Loc.353.

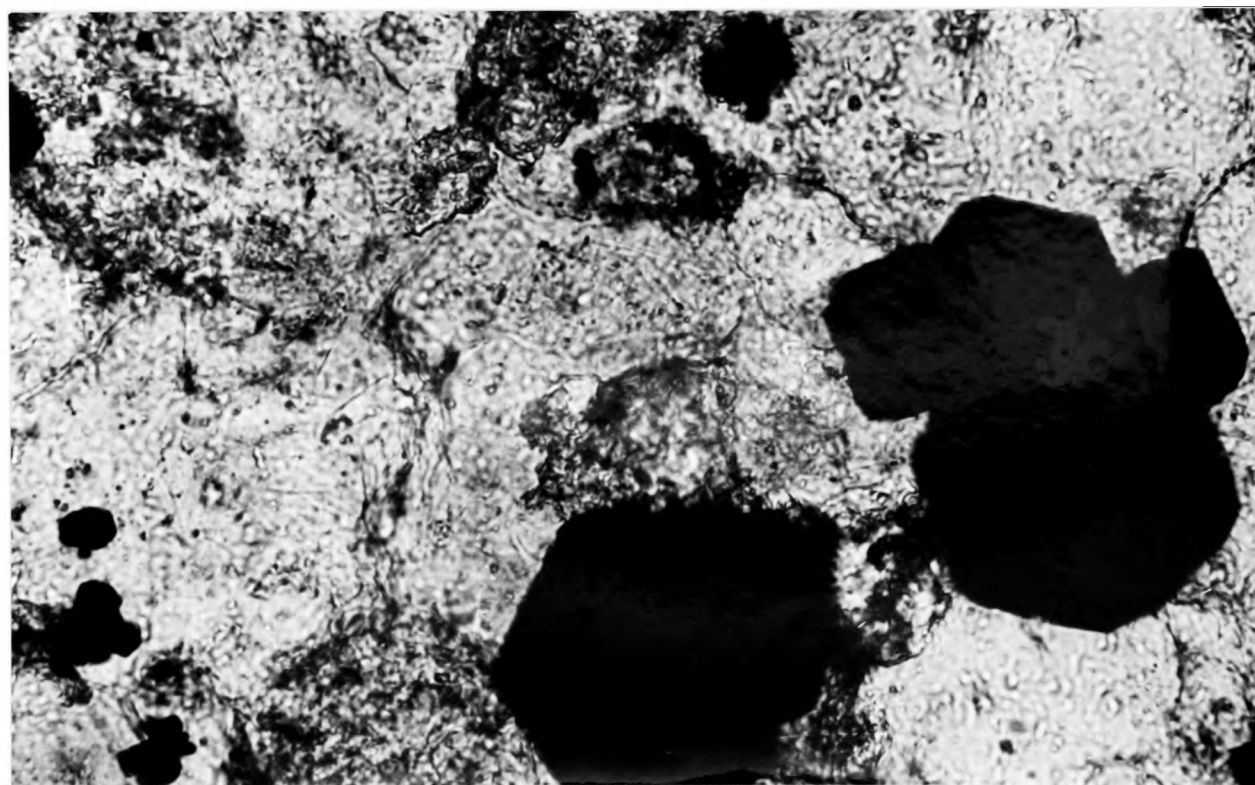


Fig. 2.4If

50 μ

Secondary purite in calcareous siltstone. Slide 050.4. Unit 8. Loc.055.

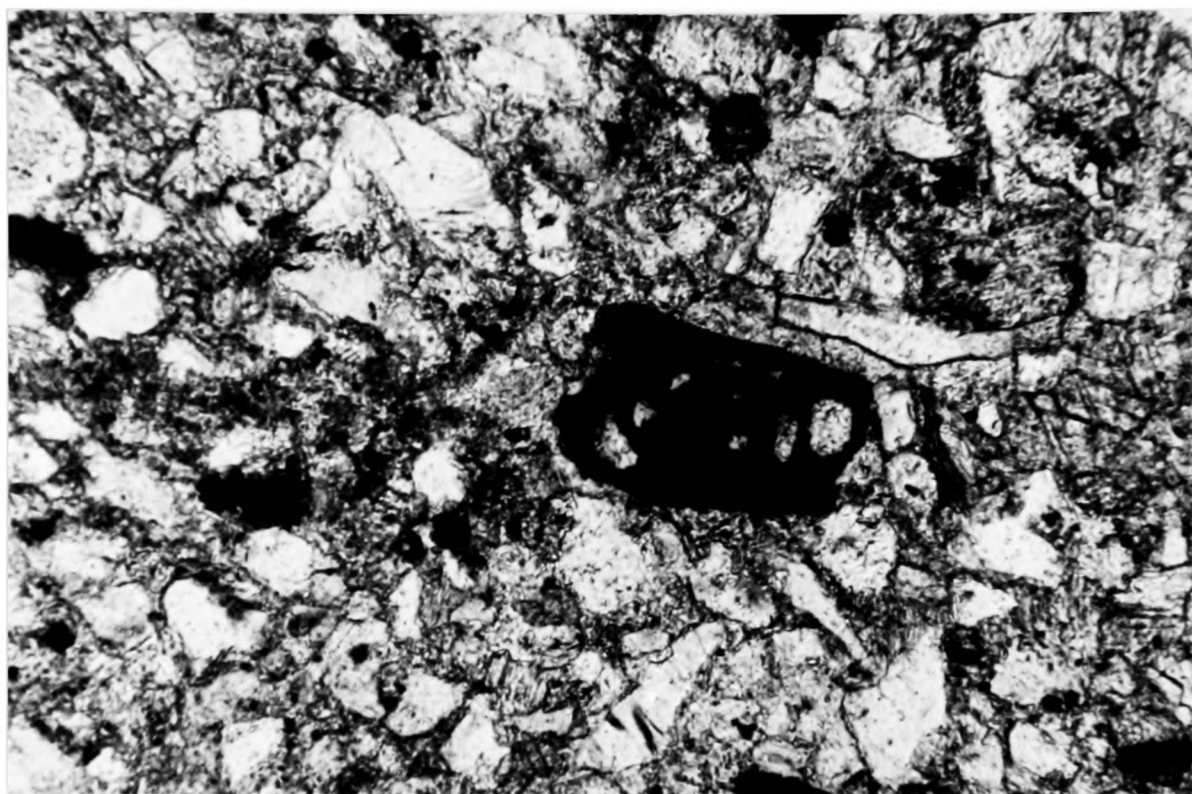


Fig. 2.4Ig

100 μ

Pyritised Endothyrid foraminifera in calcareous siltstone.
Slide 109.3a. Unit ? Loc.109.

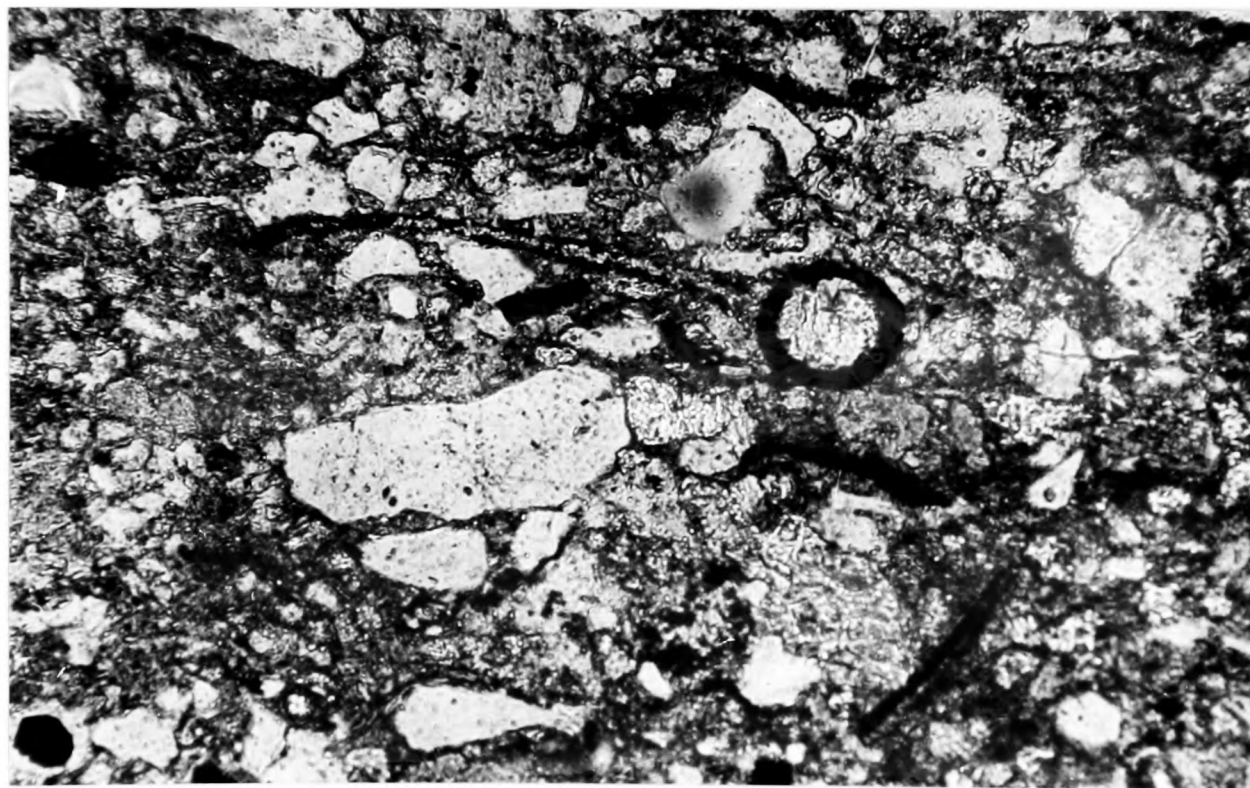
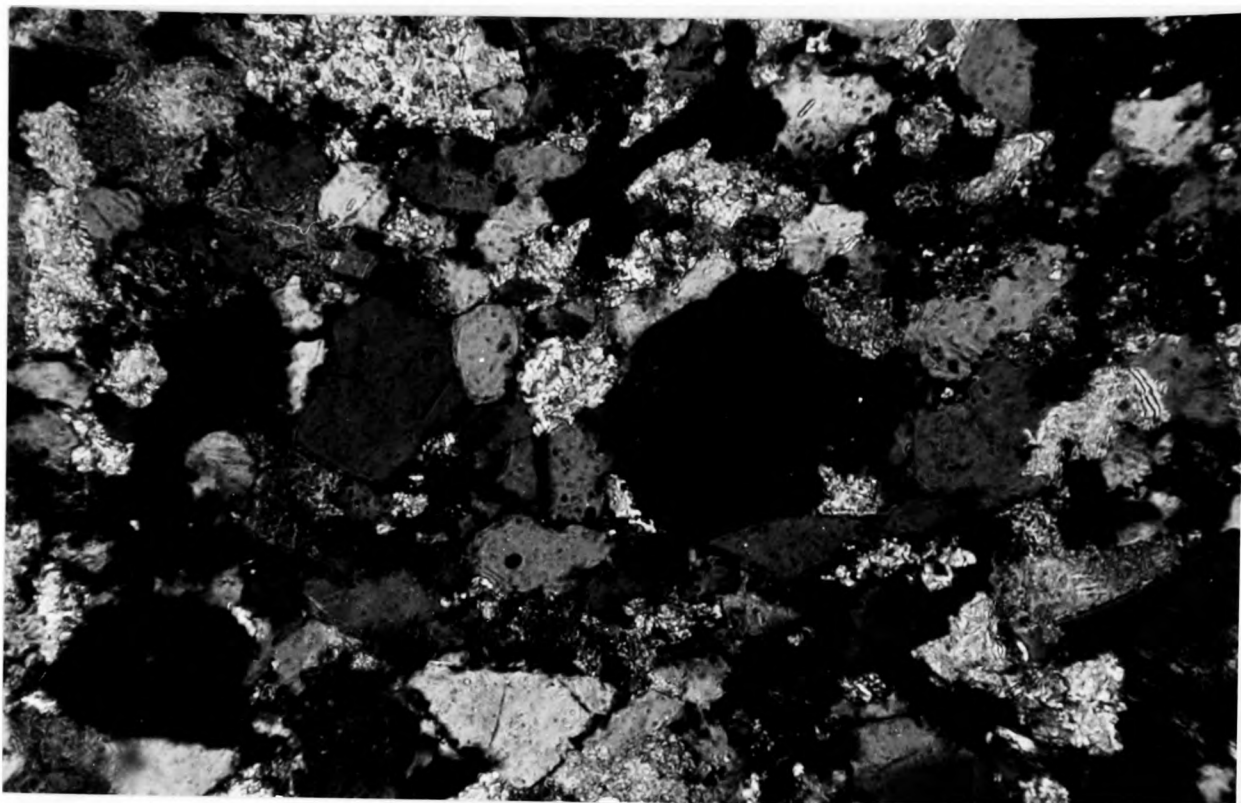


Fig. 2.4Ih

100 μ

Poorly sorted calcareous siltstone with abundant pyritised
fossil fragments. Slide 153.2a. Unit β . Loc.153.



Crossed polars.



100 μ

Typical pale calcareous quartzose siltstone.

Slide 353.12b. Unit 8. Loc.353.

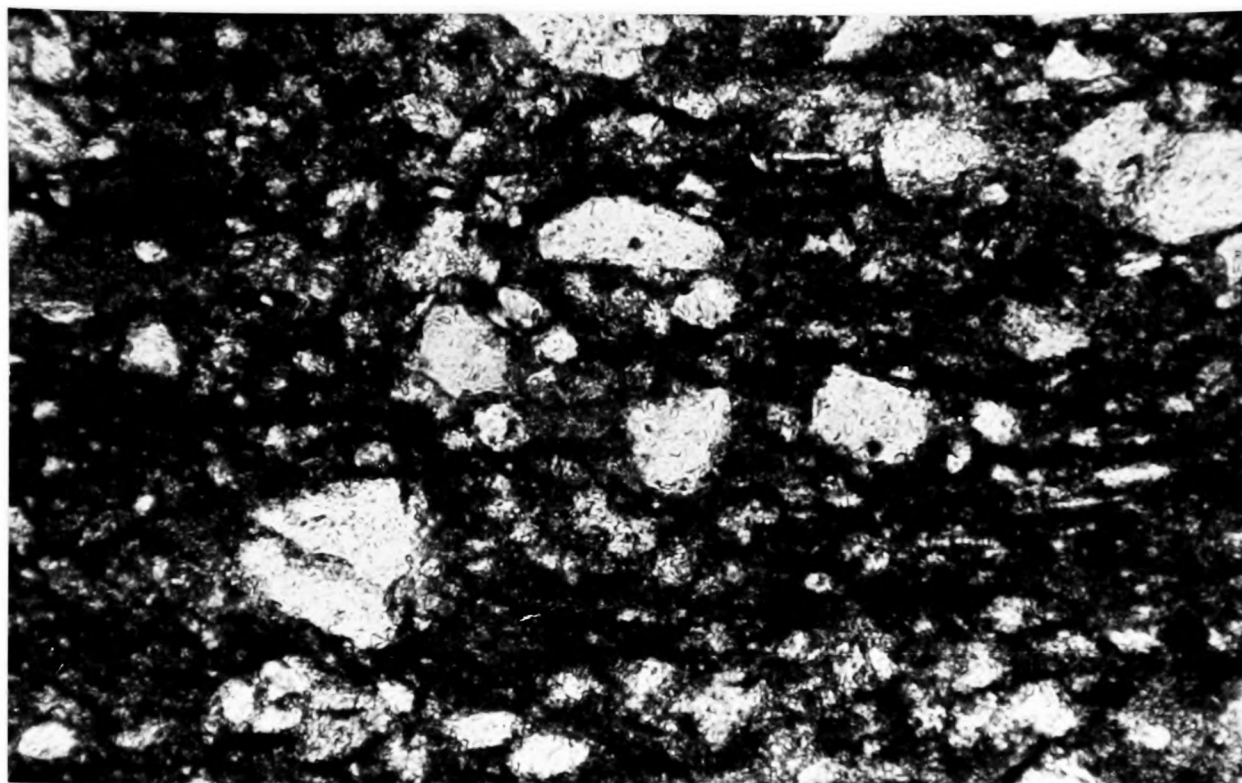


Fig. 2.4Ik

50 μ

Dark calcareous siltstone with a high proportion of matrix material. Slide 353.7. Unit γ . Loc.353.

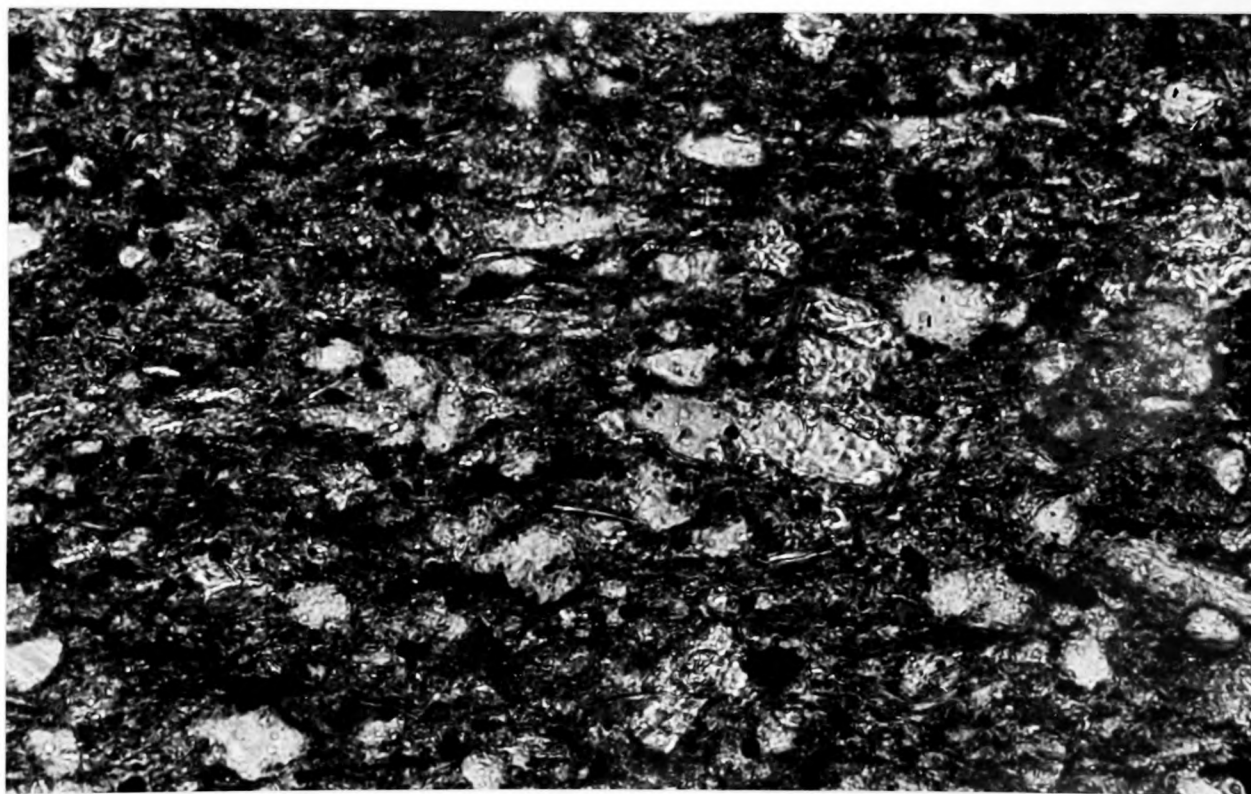


Fig. 2.4Il

100 μ

Dark calcareous silty mudstone top of calcareous siltstone bed with a few flakes of detrital mica. Slide 153.2b. Unit β . Loc.153.

Fig. 2.4Im.

	353.12A	353.12B	353.12C	353.11
	Lower Coarse laminæ partly convolute	Ripple lamin- ation above convolutions	Dark siltstone just above weak pale laminæ	Fine grained dark graded top of bed
Detrital Quarts	39.0	43.8	29.2	24.5
Secondary Quartz (Chert & rims)	12.2	10.6	5.1	-
Carbonate	20.5	39.8	12.1	-
'Matrix'	4.6	-	36.5	61.8
Kaolinite	8.6	3.2	7.6	not stained
Shale Fragments	11.4	1.4	-	-
Mica	-	-	2.3	6.6
Carbonaceous material	3.8	0.8	7.2	7.0
Pyrite	-	0.4	-	-

500 points counted for each slide.

'Matrix' - includes all fine grained material present such as clay minerals, finely divided carbonaceous material, framboidal pyrite, and probably some very fine grained carbonate.

2.4J

ORIGIN OF CALCAREOUS SILTSTONES.

Holdsworth (thesis 1963 p.226) concluded that the calcareous siltstones were the deposits of weak turbidity currents. In support of this theory are the frequently observed sequences of internal structures resembling base cut-out sequences in turbidities. The grading observed within the beds is also consistent with a turbidity current origin.

Other features of this lithofacies are not consistent with an orthodox turbidity current origin, and these features indicate that the beds were not deposited by the same kinds of currents that deposited the protoquartzite beds.

Features that are absent from the calcareous siltstones but which could be expected to occur in turbidites are:

1. Absence of complete a to e sequences.
2. Absence of a proximal area with thick deposits.
3. Complete absence of sole markings.

1. The first phase of deposition from a turbidity current gives a graded, poorly sorted division which is presumed to be deposited in the upper flow regime (Walton 1967, Walker 1965). The absence of this division in the calcareous siltstones of any part of the basin area or its surrounds indicates that the currents responsible for the deposition

of the calcareous siltstones never flowed in the upper part of the upper flow regime while in a depositional phase. It is conceivable that such currents existed but they have left no deposits, being purely erosional. But no evidence of extensive erosion has been seen at the base of any beds and it is thought unlikely that any currents existed which flowed in the upper part of the upper flow regime during deposition of the calcareous siltstones.

2. Coupled with the absence of the a division is the lack of any marked proximal area from which the calcareous siltstones could be derived. It has already been indicated that the source area of the calcareous siltstones is not known with certainty, but the petrography of the detrital minerals indicates that they were probably derived from the same land area as the protoquartzites. There is very little change in thickness of the units over the basin area. The only areal variations of importance appear to be the developments which contain fewer and thinner calcareous siltstone beds with no coarse parallel laminae and only a few ripples. These developments were seen at Sprink and Upper Shirkley in the south of the area in the region of the proximal protoquartzites. No instances of individual beds or units thickening towards the margin of the basin have been observed. This feature is in marked contrast to the protoquartzitic

turbidites where a marked thickening towards the source area has been demonstrated.

3. The complete lack of sole structures would be unusual within a turbidite bearing sequence but it must be remembered that sole structures are less frequently seen on base cut-out beds (Bouma 1962). Other reasons for the absence of sole structures are also possible. Scour structures are not usual on distal turbidites of cde type, and the absence of tool markings could be due to an absence of tools since the plant fragments and shale pellets thought to be responsible for tooling in the protoquartzites, are rarely found in the calcareous siltstones. No large plant fragments are present and only rare examples of included shale fragments have been seen (Fig.2.4Cb). Very faint tool marking could be obscured on the bases of the beds since they are usually carbonate cemented to the upper one or two millimetres of the underlying shale.

The three features described above are negative features. There are features of a positive nature which also indicate that these beds differ from turbidites.

4. Ripples with frequent truncated lamination.

5. Parallel lamination with strongly contrasting laminae.

4. The type of ripple lamination characteristic of the C division of turbidites, is Type 1 of Walker (1963). In this type deposition was continuous and truncation of ripple laminae rare. Within the calcareous siltstone

beds truncation of ripple laminae is of frequent occurrence (Fig. 2.4Cd.) and gives ample evidence that deposition was not continuous at any one place during deposition of a bed. Moving ripples were present and whilst they were migrating over the sea floor the current forming them was able to remove by sorting the dark fine grained material, so leaving the ripples relatively matrix free in contrast to the dark siltstone. During deposition of a turbidite the rapid deposition of the ripple division does not permit such marked sorting to take place.

5. The parallel lamination present in the calcareous siltstones also differs from that seen in turbidites particularly in the upper part of the bed. The pale laminae are well sorted with respect to the dark siltstone, although they contain quartz grains similar in size to those present in the surrounding dark siltstone. The pale laminae must have been deposited by some mechanism which allowed sorting to occur and which did not operate during deposition of the dark siltstone.

The bases of the pale laminae are often slightly irregular whereas the tops are usually sharp. Some mixing took place of the pale material and the underlying dark siltstone before the top of a pale lamina became stabilised. This could have occurred if the pale lamina was present as a sorted traction carpet on the sea floor. The moving traction carpet could have incorporated the dark underlying

material within its base, but when the carpet stopped moving no incorporation of material within the top of the light material would take place - giving the deposited laminae a sharp top.

Another possibility is that the pale laminae formed by winnowing of previously deposited dark siltstone to leave the pale quartz fraction as a laminae. This means erosion took place at several times within the deposition of an individual bed, and while this is not impossible, it seems more likely that the laminae were formed by sorting during deposition, rather than as a lag deposit of a minor erosive phase. The micro-loadings visible at the bases of some pale laminae also indicate that the laminae were being deposited as pale, matrix-free quartz layers, liable to sink into the underlying dark siltstone, rather than that the laminae had erosive bases.

The current that deposited the parallel laminated and rippled parts of the calcareous siltstones sometimes was able to sort out the material it was carrying giving the marked pale laminae while at other times all grain sizes were deposited together producing dark siltstone. This alternation could have been due to changes in velocity of the current or to changes from laminar to turbulent flow or both of these factors in combination. These features of sorting during deposition are absent from the proto-quartzitic turbidites, and no comparable features are known

to the author in the extensive literature on turbidites.

Isolated ripples in variable position.

Isolated ripples that occur within some beds, often associated with weak parallel lamination, are out of place in the normal sequence of structures observed in turbidites since in the siltstones they occur within the d-e divisions, (Figs. 2.4Ch,i,k,l.). This would imply speeding up of the depositing current during its final stages - which is very unlikely. These isolated ripples are possibly due to a current reworking dark, poorly sorted siltstone, but, like the pale laminae, may be due to sorting during deposition. The ripples have sharp tops against dark silty mudstone. Isolated well sorted ripples in unsorted siltstone or silty mudstone are not known from definite normal turbidite sequences. Isolated ripples do rarely occur but normally rest with an erosive contact on mudstone.

For the above reasons, both positive and negative it seems that the currents depositing the calcareous siltstones differed from the normally accepted concept of the turbidity current. The most unusual feature is the marked and frequent change in sorting of the deposit thought to result from alternations between laminar and turbulent flow. When this feature is combined with a lack of (a) sole structures, (b) the a division of a turbidite, and (c) a proximal area of deposition the calcareous siltstones contrast markedly

with the protoquartzites described, which conform to the normally accepted image of a turbidite sequence.

STRENGTH OF CURRENTS DEPOSITING CALCAREOUS SILTSTONES.

The majority of the currents responsible for deposition of the calcareous siltstones did not exceed that velocity required to produce ripples in fine sand, and were thus confined to the lower flow regime (Simons et al. 1961). A few beds show parallel lamination of a coarse type which grades up into ripple lamination. This coarse parallel lamination was produced at velocities only slightly greater than those required for rippling and represent deposits due to currents just within the lower part of the upper flow regime (Simons et al. 1961).

Holdsworth (thesis 1963 p.221) points out that since similar sized quartz grains occur in the pale laminae and the adjacent poorly sorted dark siltstone the maximum transporting powers of the currents responsible for both were similar. This adds further evidence that it was the nature, rather than the strength of the current that was responsible for the lamination produced.

The actual velocity involved in depositing the lower laminations is difficult to determine, since the type of current responsible is unknown. If the fine to medium sand was introduced by traction as bed load, velocities of

of c.40 cm/sec. would have been required (Sundborg 1956), Fig.23). However the data of Sundborg are for clear fresh water - a situation probably very different from that in the basin. Kuenen (1967 p.209) using clay rich suspensions found that a current of 50cm/sec. with a load of 160gm/litre clay would carry in suspension a load of fine sand and silt of around 55gm/litre. If the lower laminae are due to deposition in the upper flow regime velocities greater than these may have occurred. These velocities are greater than those that have been observed for normal currents in deep seas where deposition of clay takes place, (20-30cm/sec. Kuenen 1967 p.211), and this factor is a serious objection to the deposition of extensive sand beds by normal bottom currents moving their load by traction, but is in support of the conditions presumed to occur in turbidity currents.

It should be stressed that the present area is small relative to typical flysch-filled troughs and to the deep ocean areas and basins where turbidite deposition is at present taking place. Water depth was probably not excessive. Thus a comparison of bottom current velocities with such environments may not be valid, and in the area of the north Staffordshire basin stronger currents were possibly operating. The evidence is that the currents only very rarely reached velocities sufficient to bring medium sand into the area and

to deposit it at velocities greater than those required for ripples to form.

NATURE OF CURRENTS DEPOSITING CALCAREOUS SILTSTONES.

It has been shown that several features in the calcareous siltstones cannot be easily explained by the action of turbidity currents of the type that deposited the protoquartzites. There are, however, several features which are equally difficult to explain without postulating the existence of intermittent sediment laden currents flowing at velocities greater than those attained by normal bottom currents in depositional areas. The question of absolute velocity has been mentioned above. Deposition of both coarse and fine material took place at the same time during deposition of the dark matrix bearing parts of the bed. This implies deposition from an overloaded suspension current (Kuenen 1967 p.213).

Deposition of graded beds of large areal extent and even thickness cannot be satisfactorily explained by the action of normal traction currents since the transport of material at ripple velocity from one side of the present area to the other would take over a month (calculation based on Kuenen 1967 p.213) and the evidence available (Kuenen 1967 p221) is that a bed grading from parallel lamination up into ripples takes only a few

hours to be deposited. Evidence from the lower laminations favours rapid deposition, since delicate shale fragments are preserved in the bases of the beds and internal erosion is absent near the bases of these beds. The truncation of laminae observed within the ripple intervals is not usually apparent near the base of the interval, but its presence must indicate that traction of material together with local erosion and deposition was occurring during the deposition of part of the ripple interval. This last feature is in contrast with the climbing ripples usual in typical turbidites (Kuenen 1967 p.219).

The features found in the calcareous siltstones which are in accordance with an origin from turbidity currents are the basal coarse laminae with some clay matrix presumably formed by deposition in the upper flow regime, the grading of the beds, and the unsorted dark siltstone layers.

On the other hand, if the beds were deposited by overloaded suspension currents the upper parallel lamination and the ripple lamination, as well as the marked variation in sorting, are difficult to explain. The regional features of lack of a proximal deposition area, total absence of both division 'a' and sole structures are similarly unusual.

It is clear that neither the turbidity current in its classic concept (see Kuenen 1967; Dzulynski and Walton 1965), or a marine current moving its load by traction can explain all the features present.

The currents responsible for the calcareous siltstones had several properties that can be deduced from their deposits. These are as follows:-

- 1) Intermittent nature allowing carbonaceous shale and mudstone to accumulate between deposition of beds.
- 2) Maximum (inferred) depositional velocities in lower part of upper flow regime.
- 3) Ability to deposit clean, relatively well sorted, matrix-free laminae (apparently by traction).
- 4) Ability to deposit unsorted matrix-rich siltstone between the laminae of 3) above (apparently from suspension).
- 5) Ability to rework dark siltstone to produce isolated matrix-free ripples of fine sand with sharp tops against silty mudstone.
- 6) Energy to transport fine-medium sand from margin to centre of north Staffordshire basin area.

It is assumed that one type of waning current is responsible for all these features. If two were present - say a traction current affected by a turbidity

current - both would have to be intermittent (to allow shale deposition to occur) and they would have to be synchronous in action for each bed - a clearly untenable hypothesis.

No deposits such as those described here have been produced experimentally to the authors knowledge, thus no information other than that regarding the formation of individual structures is available.

The most feasible explanation is that the beds were emplaced by some form of suspension current which was acting over a longer period of time than the typical turbidity current. It was inhomogeneous in that it could transport material both in suspension and by traction and could deposit from either transporting phase. It is also capable at times of erosion to produce the isolated ripples and truncated ripple lamination. This last feature shows that it was not an 'overloaded' suspension current, which probably explains its ability to keep matrix material in suspension while depositing matrix-free laminae.

The rather dilute nature of the currents means that the density of the suspension current would be less than that of the typical turbidity current. In its initial stages of flow it would lack the absolute mass and hence the acceleration of the traditional turbidity current, hence it possibly would not reach such a high velocity

of flow and this may be reflected in the total absence of the 'a' division typical of accepted turbidities.

The conclusion is that the calcareous siltstones were deposited by intermittent inhomogeneous, dilute suspension currents which frequently carried the coarser part of their load by traction. The point or area of origin, and the mechanism by which such currents could be initiated is not known.

SECTION 2.5

SHALES INTERBEDDED WITH THE CALCAREOUS SILTSTONES.

Between individual siltstone beds the dominant lithology is shale of a black or dark brown colour. The shale is only slightly silty, quartz grains making up less than 5% of the rock. The shales are usually slightly calcareous and may contain P. corrugata. Carbonaceous material is abundant, usually occurring as black opaque or translucent brown streaks parallel to the bedding. Flakes of clay minerals are detectable in thin section but cannot be identified with certainty apart from the occasional plates of mica, which may be visibly altered to low birefringent material.

Diffraction traces of shales from between calcareous siltstones indicate that the main clay mineral present is a random mixed layer mica-montmorillonite with a basal spacing ranging from 10 -14Å with an average value of about 11.5Å. This material is probably the alteration product of detrital mica or illite in the marine environment. Kaolinite is present but only in small amounts when compared with the shales associated with the protoquartzites. The kaolinite that is present may be detrital kaolinite but is more likely to be a product of recent weathering of the mica-montmorillonite by acid ground water. Kaolinite would have been unstable

in the marine environment of the calcareous siltstones and any that was present in the original sediment was probably upgraded to illite.

It has already been suggested that the kaolinite seen in thin sections of the calcareous siltstones was of secondary origin. It is not possible to tell when the crystallization of the kaolinite took place. Kaolinite is unstable in the normal marine environment (Grim 1951 p.229, Keller 1956 p.2701) and should be preserved only if the supply of detrital kaolinite is too great to enable all of it to be converted to illitic material, or if it was formed during diagenesis as seems likely in the case of the calcareous siltstones. Degens et al. (1957 p.2443) found that the illite/kaolinite ratio for Carboniferous (Pennsylvanian) shales from the Appalachian coal basin in Pennsylvania was greater for marine than fresh water shales. Their ratios were based on areas of the 001 peaks. The marine shales described here have lower illite/kaolinite ratios than those of the shales associated with the land derived protoquartzites. Chlorite has not been detected with certainty in the shale between the calcareous siltstones.

The clay mineralogy of these shales reflects their marine origin and contrasts with the clay mineral assemblage of the protoquartzites and their associated shales (Section 2.1) which contain, in addition to mixed

layer mica-montmorillonite, chlorite, well crystallized mica, and abundant kaolinite which do not reflect the nature of the depositional environment but that of the source environment. Deposition was probably too fast to allow them to adjust to the depositional environment.

The shales between the siltstone beds usually grade downwards into the silty mudstone tops of the calcareous siltstone beds. The clay minerals in the silty mudstone are identical to those in the shales with the exception that well crystallized 10\AA mica is usually present and gives a separate peak on the diffractometer trace. It is probable that the larger grain size of the mica flakes in the siltstones prevented them being totally converted to mixed layer mica-montmorillonite.

SECTION 2.6

MARINE BAND SEDIMENTS.

All the marine bands present within the part of the succession studied consist mainly of dark brown to black fissile shale. Fossils are all crushed with the exception of crinoid stems which are either preserved uncrushed or are present as voids from which the carbonate has been removed after compaction of shale.

The C. malhamense and E. grassingtonense bands occur in identical situations within the successions and can be considered together. They consist of dark fissile shale which overlies unfossiliferous dark shale mudstone. The dark fissile shale contains the crushed fauna of the band. The marine shales are succeeded upwards by calcareous siltstones and the faunas of the two marine bands are found in the shales between the few lowest siltstones in Units β and γ . In the E. grassingtonense band brown weathering bands are present up to 2cm. thick. The bands contain less carbonaceous material than the dark fissile shales and relict spicular structures and radiolaria occur in them. It is probable that these bands represent very weak beds of 'calcareous siltstone' type and were deposited by weak currents.

The E. erinense E2a2 band is the thickest goniatite bearing band in the succession and is about 5 metres thick (Log. fig.1.3i). A fauna is found throughout

the band with the exception of a 10cm. band 1.2 metres above the base of the band which may only be a local feature of the Croker Hill area. Thick shelled goniatites are most frequent about a metre from the base of the band, but can be found right up to within 0.5m. of the top of the band. In the top 2m. of the band brown beds up to 2 cm. thick are seen similar to those in the E. grassingtonense band and appear to represent very weak 'calcareous siltstone' beds.

In the middle of the band the shales are often cemented by secondary carbonate. Some calcite is present but ankerite or dolomite is probably more abundant. The cementing carbonate destroys the fissillity of the shales and converts them to platy beds a few centimetres thick.

Within the band a few scattered calcite concretions (bullions) occur (Fig. 2.6a). Within the concretions uncrushed fossils such as goniatites indicate that the concretions formed early in diagenesis before compaction of the shales. Measurements of equivalent laminae lateral to bullions shows that the shales have been compacted considerably and represent at most only 0.2-0.3 of the original thickness of deposited sediment, (Fig. 2.6b).

Fig. 2.6b.

Thickness of Bullion mms.	Shale Equivalent of Bullion mms.	Compaction Ratio.
69	20	3.4
64	18	3.6
51	15	3.3
71	20	3.5
66	13	5.2

A bullion from Hurdlow Loc.009(Fig. 2.6c) shows one of the thin brown beds preserved in an uncompacted form and this confirms that they were introduced by weak currents. Small uncrushed goniatites, lamellibranch shells, and carbonaceous material which are common in the dark parts of the bullion, are absent from the brown band. The brown bed contains only small radiolaria and spicular structures which are scattered within the band. All the material of the brown band could be transported by a weak current whereas the fossils and carbonaceous material of the rest of the bullion are typical of quiet pelagic deposition. Associated sponge spicules are found in the dark parts of the bullion indicating very quiet conditions of deposition. The material forming the brown bands appears to have been introduced mainly in suspension by very weak currents which failed to produce any fine lamination within the bed.

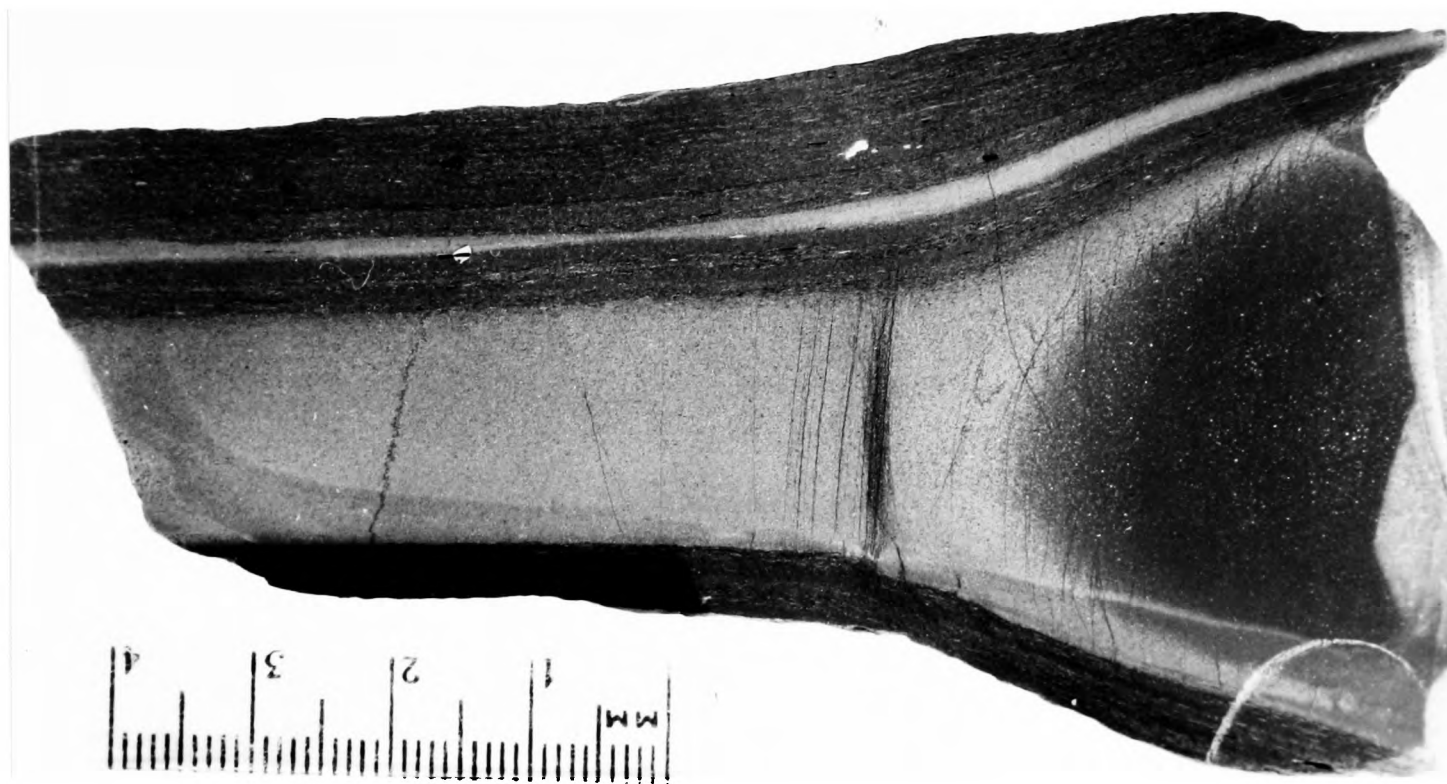


Fig. 2.6a

Thin silty bed in marine band shale. The silty bed passes laterally into an early diagenetic calcareous concretion (bullion) at right of photo. E2a2 Marine Band. Loc.351.

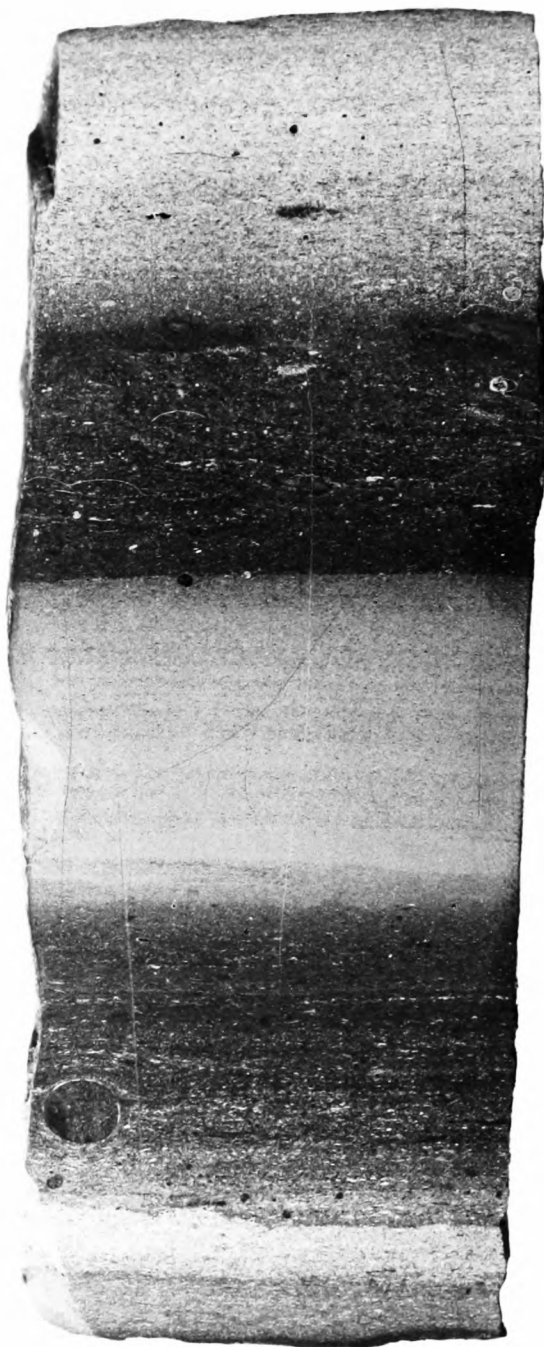


Fig. 2.6c

Etched section of bullion showing pelagic deposits with numerous organic fragments, and a pale fine grained bed in the centre of the bullion with a sharp top and gradational base. The pale bed contains no fossil fragments and has weak parallel lamination and was probably deposited from suspension by a weak current. E2a2 Faunal Band. Loc.009.

The E. erinense E2a2 faunal band does not pass upwards into calcareous siltstones as do the two lower bands but the thin brown slightly silty horizons within it are very similar to weak calcareous siltstone horizons and appear to be formed in the same way. Thus, although it appears at first sight that the marine band is followed immediately by protoquartzites in contrast to the underlying horizons, in reality the thick marine band contains weak 'calcareous siltstone' type beds and there is an interval of 70cm. of unfossiliferous shale-mudstone above the faunal band before the first protoquartzitic bed is seen. The E2a2 faunal band can thus be likened to a very weak development of calcareous siltstones with a goniatite fauna throughout whereas the lower two calcareous siltstone units have goniatite bearing horizons confined to the base of the unit, the rest of which contains only a P. corrugata fauna.

Above Unit C of the protoquartzitic turbidites the other marine bands described consist of shale and shale-mudstone which is sometimes calcareous. The horizons are all of similar sediment types apart from slight variations in fissility of the shale and silt content which have not been investigated in detail.

The Ct. edalensis E2b1 faunal band is of grey to buff weathering shale with a smooth texture which is

not seen in the other bands. This band also contains calcitic bullions which are rounder than those in the E2a2 band but lack of any distinctive laminations makes determination of compaction ratios impossible. The fossils within the bullions are uncrushed and so the bullions formed early in diagenesis before compaction took place. Bullions are also present in the upper C. cf. C. subplicatum horizon where they occur on a specific horizon and frequently contain wood fragments as well as goniatites which are concentrated in a bed through the centre of each bullion. In this case the concretion may have nucleated on the organic remains but in many cases no specific object is found in a bullion on which it has obviously nucleated.

The bullion limestones have been described by Holdsworth (1966) who noted that those in E2b1 - E2b2 tended to contain a high proportion of detrital material and few radiolaria in comparison to bullions higher in the succession. This is generally confirmed by the present work although bullions have now been found in the E2a2 band and one only in the E2a1 band. Radiolaria are rare in these bullions and are always of the spumelline type.

Petrography of Marine Shales.

The only detrital minerals that can be identified with certainty are quartz and mica. The quartz grains are scarce and always of silt grade or finer. They never constitute more than 5% of the shale. Flakes of micaceous material are frequent and some are obviously detrital in nature and represent muscovite. The majority have low birefringence and may be diagenetic in origin or could equally be degraded detrital material.

Carbonaceous material is abundant usually forming black opaque streaks parallel to the bedding. There is also a considerable amount of translucent brown organic material present. Secondary carbonate is frequent and often occurs as small spherical bodies with fibrous structure. Shell material is usually preserved as carbonate. Fine grained clay material cannot be identified optically. Pyrite is present in the shale both as minute well crystallised cubes and as spherical framboids. It is secondary in origin and, as in the calcareous siltstones, is frequently associated with fossil fragments.

Clay Minerals in Marine Band Shales.

Diffraction traces of whole rock specimens of the marine band shales show them to be similar to the shales found interbedded with the calcareous siltstones

(Section 2.5). Kaolinite is present in only small amounts and in several specimens (from E1c, E2a1, E2b1 bands) is absent or below the detection limit of the method. The main clay present is an irregular mixed layer clay with basal spacings ranging from 10-14Å, probably a mica-montmorillonite. Only occasionally was a marked peak obtained for 10Å mica and this was always lower than the broad peak of the mixed layer material. Chlorite could not be detected in any samples of the marine shales.

The fact that the marine shales have only one important clay type present - the mica-montmorillonite - would seem to indicate that this was the stable clay material of the environment (or its present day alteration product). The slow deposition of the marine bands and the very fine grained nature of their constituents are both factors which would have helped in bringing the clay minerals into equilibrium with the marine environment before burial took place. This is in marked contrast with the coarser and rapidly deposited protoquartzitic turbidites where kaolinite, mica, mica-montmorillonite and chlorite can all be easily detected.

Conclusions.

The marine band sediments were deposited slowly and are highly compacted as shown by the bullions. The

faunal concentration within the bands is possibly due more to the slow deposition and conditions favouring preservation than a great absolute abundance of individuals. The bullion textures indicate that the original sediment was highly water charged, but in contrast to the bullions described by Holdsworth (1966) the sediment contained few radiolaria, and has evidence of weak current activity. The sediment surface was also capable of supporting a benthonic fauna especially in the E2a2 faunal band.

The main clay mineral present is a mixed layer mica-montmorillonite which was probably the stable form in the marine environment. The diagenetic minerals are calcite and dolomite-ankerite together with pyrite. Siderite is unknown within the marine bands studied.

PART 3.

LOCALITIES OUTSIDE THE NORTH STAFFORDSHIRE BASIN AREA.

3.1

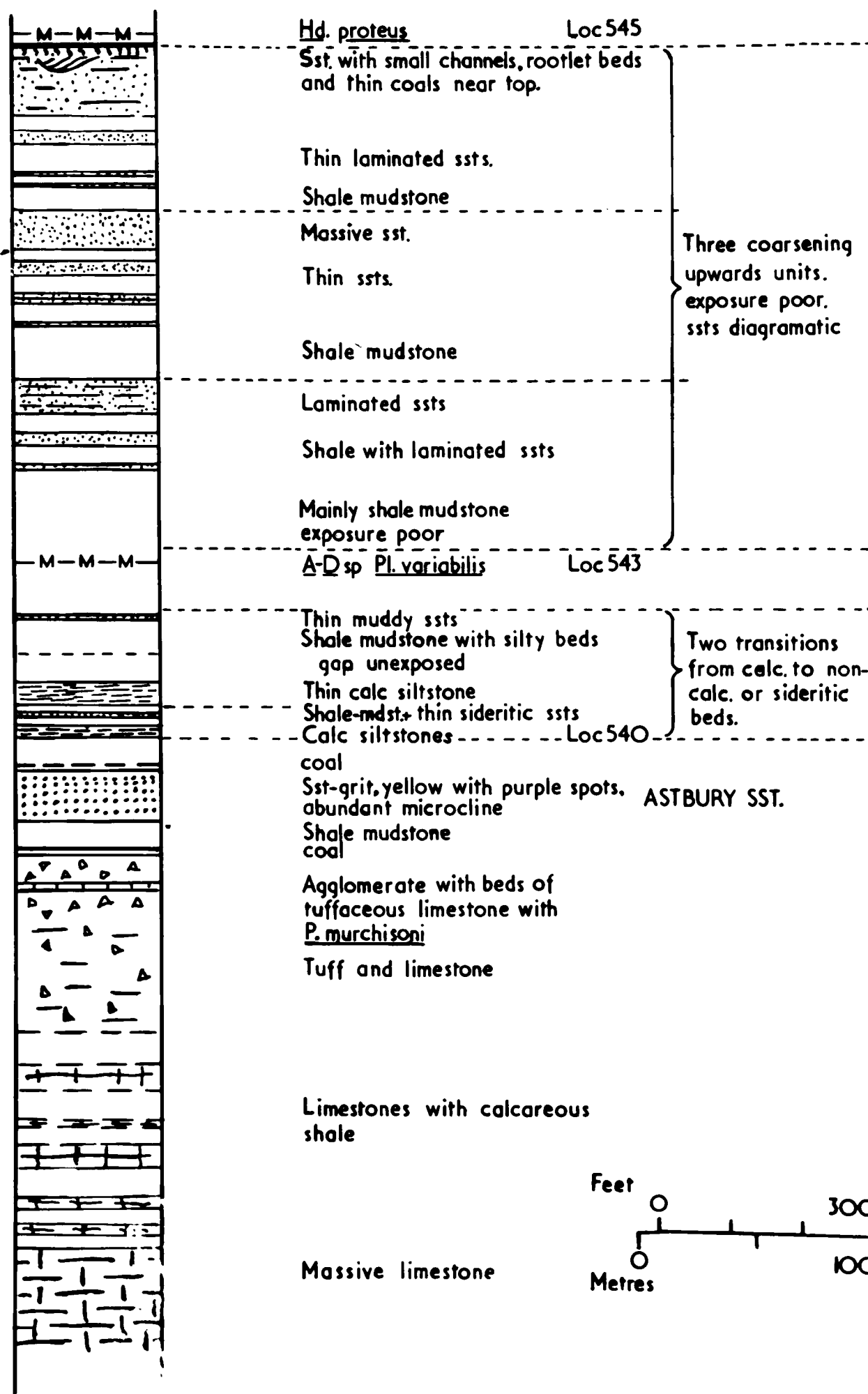
ASTBURY LIME WORKS.

The section exposed from the Viséan limestones, seen in the disused and water filled quarry of the Astbury Lime Works, up as far as the Hudsonoceras proteus band of the Namurian, exposed in the gannister quarry (Loc. 545), is totally unlike that seen elsewhere in the area. The section as measured by the author is presented in Fig.3.1a. The thicknesses of the lower parts of the section in the Viséan are taken from Gibson and Hind (1899), but the general succession can still be verified although exposures are now very poor.

The massive limestones at the base of the succession are not now exposed, being under water. The fauna that was obtained from them is unusual in that it contains elements of the Carboniferous 'Reef Fauna' (cf. Wolfenden 1958) Schizophoria resupinata (Martin); Spirifer striatus (Martin); S. trigonalis Martin; Pugnax acuminatus (J. Sow); together with species of Antiquatonia, Buxtonia, Echinoconchus, Eomarginifera, Linoproductus and Gigantoproductus. This indicates that during Viséan time the fauna of the area was of 'Reef' or 'Massif'

LOG. ASTBURY SECTION.

Fig 3.1a.



type rather than that of the 'Basin' type and hence the limestone is probably also of 'Reef' or 'Massif' type rather than of 'Basin' type with bedded dark limestones and shales. Ramsbottom (in Evans et al. 1968 p.10) considers the fauna in the Astbury Limestone to be D1 in age.

The calcareous shales with limestones that overlies the massive limestone are not now exposed in the quarry. The igneous rock present in the quarry consists mainly of basic tuff and agglomerate, but there may be a vent present in the northern end of the quarry although conclusive evidence is now lacking. The tuff and agglomerate is green in colour and weathers brown. Numerous fragments of vesicular lava are present. Impure limestone bands within the tuffs (Loc. 544) contain corals which include Palaesmilium murchisoni, normally a D1 species. This record possibly indicates that at least part of the Astbury Limestone-Shales of Evans et al. (1968 p.9) are D1 in age.

The tuffs and agglomerates are overlain by dark shales and mudstones which are sometimes reddish in colour. A thin coal seam was reported from near the base of these shales by Gibson and Hind (1899), and there is another present at the top immediately below the feldspathic sandstone known as the Astbury Sandstone (W.B. Evans pers. comm. 1967).

The Astbury Sandstone (Loc. 541) is usually yellow colour and is often spotted purple. Its grain size ranges from fine sand up to granule size (Pettijohn 1957 p.19) but the most distinctive feature is the abundance of fresh microcline feldspar that it contains. No comparable material is known from anywhere in the Viséan or sub-R1c Namurian to the east and its source is unknown.

Above the Astbury Sandstone a coal seam is again present and on spore evidence it is probably P2 in age (Evans et al. 1968 p.14). It seems that the upper Viséan of the area around Astbury was characterised by very shallow water, possibly with periodic emergence to give periods of coal formation. (The coal is not now exposed and there is no conclusive evidence as to whether the coal is drifted or in situ.)

Above the coal on top of the Astbury Sandstone lie about 300 metres of strata containing no marine horizons that can be assigned to a definite level within the Carboniferous. The Hd. proteus band (H2) is the next horizon that can be dated with certainty.

The exposure of the beds is poor above the Astbury Sandstone, but in Limekiln Brook at Loc. 540 some evenly bedded dark siltstones are present which are slightly calcareous in a few beds. The beds are laminated and one bed at the top of the development contains frequent

shell fragments. The fragments are mostly indeterminate but one brachiopod cross section was found in a thin section. The only macro-fossil obtained from these beds was an indeterminate orthocone nautiloid of no stratigraphic value.

Mineralogically these beds are similar to the calcareous siltstones of the basin area, containing no K-feldspar and only traces of plagioclase. Carbonaceous material is more abundant than in the calcareous siltstones of the basin area. On lithological grounds the base of these beds gives the best correlation with the base of the Namurian of the basin area where Cravenoceras cf. leion occurs (Loc. 260) at the base of the Gun Hill Siltstones (E1a-E1b) which are over 150 metres thick. The dark siltstones at Loc. 540 are only about 5 metres thick, and are followed by 9 metres of shale-mudstone with a few thin sideritic sandstones. The sandstones are only a few centimetres thick and are rather muddy. They do not have well defined upper or lower contacts with shale-mudstone nor do they show grading. Correlation such as this on lithological grounds must be treated with extreme caution, but in the absence of the relevant faunas no better correlation can be attempted. The dark calcareous siltstone lithology then returns (Loc. 542) and is also 9 metres thick. One small Lingula was collected from this development but no other fossils

were seen. After an unexposed portion of stream representing some 13 metres of strata, about 22 metres of strata are seen with grey shale-mudstone containing some silty laminations predominate in the lower half which passes up into similar shale-mudstone but with thin (1-2cm.) beds of muddy sandstone frequent near the top.

Thus in 60 metres of strata two alternations are seen of dark siltstones containing a poor marine fauna with shale-mudstones, sometimes sideritic, and containing sandstone beds. These alternations bring to mind the alternations of calcareous siltstone and protoquartzite lithofacies seen within the basin area in E2 and described previously.

Following these alternations the strata are obscured for the next 19 metres of the succession and the next exposure seen, at Loc. 543, yields the best marine fauna known to the author between the Viséan and the Hd. proteus band in this section. The band is fossiliferous through some 4 metres of strata. The basal 120 cms. consist of shales with decalcified limestones up to 10cm. thick. The fauna is largely fragmentary and in the decalcified layers it is partly uncrushed.

Faunal List. Loc. 543 (Loc. 22 of Evans et al. 1968)

Anthracoceras - Dimorphoceras fragments

Orthocone Nautiloid

Posidoniella aff. variabilis

Gastropods

Lamellibranch (indet)

Discussion.

The thin shelled goniatite fragments are too poor for identification and give no guide as to horizon. Other elements of the fauna are benthonic (Gastropod and the indet. Lamellibranch) and the remaining species, Posidoniella aff. variabilis was probably also benthonic. This last named species may give some clue as to horizon since in the basin area it does not make its appearance until the topmost band of E2a is reached and is most common in the succeeding E2b subzone. This unusual fauna is thus very tentatively placed as being E2b or younger in age. If this interpretation is correct then the whole of E1 and E2a are represented by only 90 metres of strata.

Evans et al. (1968 Fig.6) place this fauna in E1 apparently on the basis of lithological correlation of the overlying sandstones with the Minn Beds. It is perhaps significant that their record of Pl. variabilis, which they found about 17m. higher than the fauna of Loc. 543 in their 'Lask Edge Shales' is the only record of this species given by them from below the Churnet Shales. The base of the Churnet Shales being taken at the Ct. edalensis band.

Between this fauna and that of Hd. proteus about 250 metres of sediment are present. This can be divided roughly into three coarsening upwards units which become richer in sand towards the top of the succession.

The lowest unit culminates in laminated sandstones which are very poorly exposed. The second is similar but has a strong development of sandstone at its top in thick massive beds but without any visible internal structure such as cross bedding. The shales above this sandstone are exposed but no marine fauna could be detected. The third unit has thin laminated beds near the base, which pass up into thick structureless sandstone beds which are overlain by the irregular gannister beds seen in the quarry at Loc. 545. In the quarry cross-bedding can be detected in the sandstones, usually in the fill of small channels cut through massive or laminated sands. Rootlet beds are of frequent occurrence and thin coals are also present. This top unit thus resembles a typical 'Coal Measures Cycle'.

The sandstones are overlain by the marine band which contains a good benthonic fauna in addition to the goniatites. Chonetes, Schizophoria, a productid and Orbiculoidea are all present. The marine incursion in fact reworked the topmost sandstone bed since it contains both rootlets and, in its top few centimetres, impressions and moulds of orthid brachiopods.

Evans et al. (1968 Fig.6) consider that some of the sandstones present are the equivalent of the Minn Beds of E2 and upper E1 age and therefore include Units A, B and C of the protoquartzites. The sedimentology of these beds as described is totally different from those of the basin area to the east and the writer prefers to correlate the alternations of calcareous and sideritic strata seen near the presumed base of the Namurian with the similar alternations described from E1c to E2b in the basin area. The only available palaeontological evidence is the presence of P1. variabilis already discussed.

This brief summary of the succession in this area indicates that during P2-H2 times whenever a fauna is found it contains an admixture of shallow water benthonic forms which are not seen in the basin area. The presence of coal seams in P2 and again in H2 indicates that sufficient shallowing took place for emergence to occur at times.

The sediments developed reflect this situation in that the turbidites present in the basin are totally absent and appear to be represented, if at all, by thin muddy sandstones. To the east the very thick proximal turbidite developments of Sprink and Shirkley are only two miles away. Thus it would seem that during E2 there must have been a rapid shallowing of the water depth from the area of proximal turbidites to Astbury.

The absence of the marine horizons seen in the basin is thought to be real and not just due to poor exposure. If this is the case then there must have been a fundamental difference in conditions to those in the basin where the marine bands are so well developed. It is possible that marine bands were not deposited or were eroded before deposition of the next part of the succession.

Although there is no evidence of unconformity or nonsequence within the succession it would be quite easy for a break to be concealed, and the presence of coals indicates that periodic emergence did take place. The position of the Namurian succession over an area of massif or reef facies limestone possibly indicates that this area was part of the original western side of a basin area of deposition present from as early as D1 times. Periods of non deposition on such a 'high' marginal area would be very difficult to detect.

The presence of coal indicates that land was probably not far distant from the area and influx of fresh water into this shallow near shore area may have been responsible for the general non-development of the marine bands. The presence of Lingula in the dark siltstones of the succession could also indicate brackish conditions.

The succession is interpreted as representing a

generally shallow water development on the western margin of the basin area, away from the influence of the basin sediments. The non-development of marine horizons present in the basin is possibly due to influx of fresh water into this shallow water area so inhibiting the development of true marine conditions.

3.2

EDALE, DERBYSHIRE.

Jackson (1927) and Hudson and Cotton (1945) have described the faunal succession from Edale and the author's correlation of the Staffordshire succession with that of Edale has already been discussed in the palaeontological section of this thesis. Correlation by K-bentonites has also been discussed previously (Section 1.4). The sediments which occur between the marine bands in Edale have received little attention in the past and form part of the Edale Shales.

The general succession of the lithofacies within the part of the succession being studied has been shown in Fig.1.1b, and they will now be considered in relation to their lateral equivalents in the Staffordshire basin area.

Equivalents of the Protoquartzitic Turbidites.

The three units of turbidites traced over the Staffordshire basin area are absent in Edale and they are represented mainly by black shale-mudstone. Units B and C of Staffordshire are represented by 30 and 51 metres respectively of black shale-mudstone with beds of siderite up to 8cm. thick and rows of elongate nodules of siderite 5-10cm. thick. Thus the lithofacies displayed is identical to that seen in Staffordshire

in the Upper Dove Valley and at Waterhouses, localities which were beyond the reach of turbidite deposition.

Unit A of Staffordshire is represented in Edale by 18m. of dark shale-mudstone containing two thin (3cm.) beds which are calcitic and also contain pyrite nodules. This is in contrast with the siderite found in the unit in Staffordshire and is the only lateral equivalent of protoquartzite deposition in which calcite has been recognised.

The unit of shale-mudstone with siderite nodules present beneath the E2a2 faunal band in Staffordshire is also present in Edale and is 7 metres thick. This is similar to its Staffordshire equivalent which usually measures 5-15 metres.

Equivalents of the Calcareous Siltstones.

The units of calcareous siltstones described from Staffordshire are all present in the Edale succession in exactly the same relation to the faunal bands. As in Staffordshire there is no marked difference between the three units exposed and they may all be described together. The individual beds are generally less than 10cm. thick and usually only display beds showing pale quartz rich laminations up to a few millimetres thick set in dark calcareous siltstone. Occasional beds show evidence of current activity strong enough to produce ripples and one bed (Fig.2.4Cp) showed a coarser parallel

lamination underlying the rippled part of the bed, possibly indicating that the lower part of the bed was deposited in the upper flow regime. This bed also contains numerous shale fragments presumably eroded from the substrate by the current.

The thicknesses of Units α and β are c.90 and 6 metres respectively in Edale which is less than the thicknesses observed in north Staffordshire (Fig.5.1b) where Unit α , the Gun Hill Siltstones Formation, is over 150 metres thick and Unit β is 10-14 metres thick. Unit γ is 8 metres thick in Edale and thus is comparable to the thicknesses observed in north Staffordshire. The siltstones are often highly calcareous and generally appear to be richer in calcite than those of the Staffordshire area, but no quantitative work has been done to prove this point.

The Faunal Bands.

Palaeontologically all the faunal bands found in Edale have been correlated with those in Staffordshire, with the exception of the E. pseudobilingue band which is unknown in the Staffordshire basin area. This is the lowest faunal band exposed in Edale and it consists of two rows of calcitic bullions 40cm. apart set in the lowest unit of calcareous siltstones seen in the succession.

The C. malhamense band (E1c), Locs. 84, 87, also occurs within the base of a calcareous siltstone unit and fossils are most numerous in shale seams between the calcareous beds. This is a very similar situation to Staffordshire but there are more calcareous beds associated with the fauna in Edale.

The E. grassingtonense (E2a1) band ("E. aff. pseudobilingue" (E1d) of Hudson and Cotton 1943, 1945) appears to be identical with its representative in Staffordshire with the exception that a single calcitic bullion was noticed in it at Loc. 74. The E. erinense-E. ferrimontanum E2a2 band (E. bisulcatum E2a Hudson and Cotton 1945) is again similar to that seen in Staffordshire but markedly more calcitic. This feature is most noticeable at Loc. 112a where semi-continuous beds of bullion limestone are present which have not been seen in outcrops in Staffordshire where only a few small calcitic bullions are found. The L. longirostris and Ct. edalensis bands are identical to those in Staffordshire.

General Considerations.

The very marked similarity of both the faunal and lithological successions of Staffordshire and Edale is immediately noticeable. The presence of the same marine bands is to be expected since these bands occur over most of Europe, but the identical alternation of

calcitic and siderite bearing units, with one minor exception already mentioned, is surprising. This would seem to indicate that the two areas were covered by part of the same mass of water and subject to the same chemical variations. It is probable that the two areas were connected both around the west side, and over the top of the drowned Viséan massif area, the Edale exposures being just north of the Castleton reef belt.

The protoquartzitic turbidites that flowed northwards in Staffordshire along the west side of the Viséan massif did not reach the Edale area and thus must have died out in the 15 miles between the Croker Hill area and Edale. Possibly they did not reach Edale because they were unable to flow round the edge of the massif to that area, or they may have been stopped by a submarine elevation north of Croker Hill.

The conditions associated with the protoquartzites which favoured siderite deposition in Staffordshire are seen in Edale and it is thought that the siderite deposition in Edale is due to the influence of the land derived material from the south. The deposition of siderite rather than calcite or dolomite type carbonates and pyrite is thought to be due to the concentration of sulphate reducing bacteria in the sediment (which is controlled by the rate of sedimentation) and also the greater concentration of dissolved iron in the land

derived material (see Part 5). The only exception to this is the equivalent of Unit A which is slightly calcitic in Edale. This first influx of protoquartzitic turbidites is the weakest in Staffordshire and probably had the least areal effect on sedimentation. While the material that did reach Edale had the effect of reducing the amount of calcite deposited, diagenetic conditions did not alter sufficiently for siderite rather than pyrite to crystallize.

The unit of shale-mudstone with siderite nodules below the E2a2 faunal band is present in both Edale and Staffordshire but in neither area does it contain proto-quartzites. It seems probable that somewhere land derived protoquartzitic material was being introduced to the body of water but not in the areas studied where the deposits are fine grained.

Conclusions.

The close lithological and palaeontological similarities of the Edale and Staffordshire areas indicate that they were influenced by the same diagenetic conditions with the concentration of dissolved sulphur species produced by sulphate reducing bacteria controlling the diagenetic mineralogy of the sediment. Units of calcareous siltstones are common to both areas, and phases of land derived protoquartzite deposition in

Staffordshire are marked in Edale by deposition of shale-mudstone with siderite, the only exception being the equivalent of Unit A. The slightly calcitic nature of the equivalent of Unit A in Edale reflects the generally more calcitic nature of the succession in this area which was noted in the case of the calcareous siltstones and the marine bands. This is probably a reflection of its greater distance from land and the influence of the deltaic land derived material which typically has associated diagenetic siderite.

P A R T 4.

AREAL SITUATION OF THE NORTH STAFFORDSHIRE
BASIN DURING THE LOWER NAMURIAN.

The north Staffordshire basin area of deposition is briefly reviewed and is discussed in relation to the surrounding areas with emphasis on the Pendleian and Lower Arnsbergian strata. The evidence is summarised on Fig.4a*.

The succession in north Staffordshire in the upper Viséan and lower Namurian is typical of the 'Basin' or 'Gulf' facies in being thick, (P1-E2 = c510m. at Gun Hill. Hudson 1945) and consisting mainly of interbedded limestones and shales in P1 (Mixon Limestone-shales) limestones and shelly sandstones in P2 (Onecote Ssts), followed by calcareous siltstone and shales in E1 (Gun Hill Siltstones Formation or Lask Edge Shales) and protoquartzitic turbidite sandstones (Thorncliff Sandstones or Minn Beds) and shales in E2. Within the Namurian part of this succession marked areal changes have been described connected principally with the protoquartzitic turbidites. The turbidites originated to the south of the area and are thickest in the most southerly exposures seen on Lask Edge. In this area the three units of protoquartzitic turbidites and the interbedded calcareous siltstone units probably total

* Figure in folder.

over 220 metres. They thin to the north along the main transport direction to Croker Hill, and at the eastern margin of the area die out completely as the unconformity of E1 and E2 with the reef margin of the Viséan massif limestone area is approached. The main thickness development in the area appears to be along the S-N line from Endon through Lask Edge and Gun Hill to Croker Hill. To the west of this line the only exposure available at Astbury contains a development of P1 and P2 with coals, and a presumed E1-E2 succession without turbidites and without the normally well developed marine bands. The E1-E2 succession here is probably only 90 metres thick whereas at Gun Hill 7 miles to the east it is at least 270 metres and on Lask Edge less than four miles to the east of Astbury. E1 and E2 are probably much thicker in view of the nature of the protoquartzite units in E2. There is thus an abrupt change at the western margin of the area which must have involved a considerably greater amount of relative subsidence for the area to the east of Astbury. It is possible that the differential movement was controlled by faults bounding the western side of the basin but there is no direct evidence available to prove this point. It is certain that the strong turbidity currents which flowed north in the Lask Edge area were unable to influence

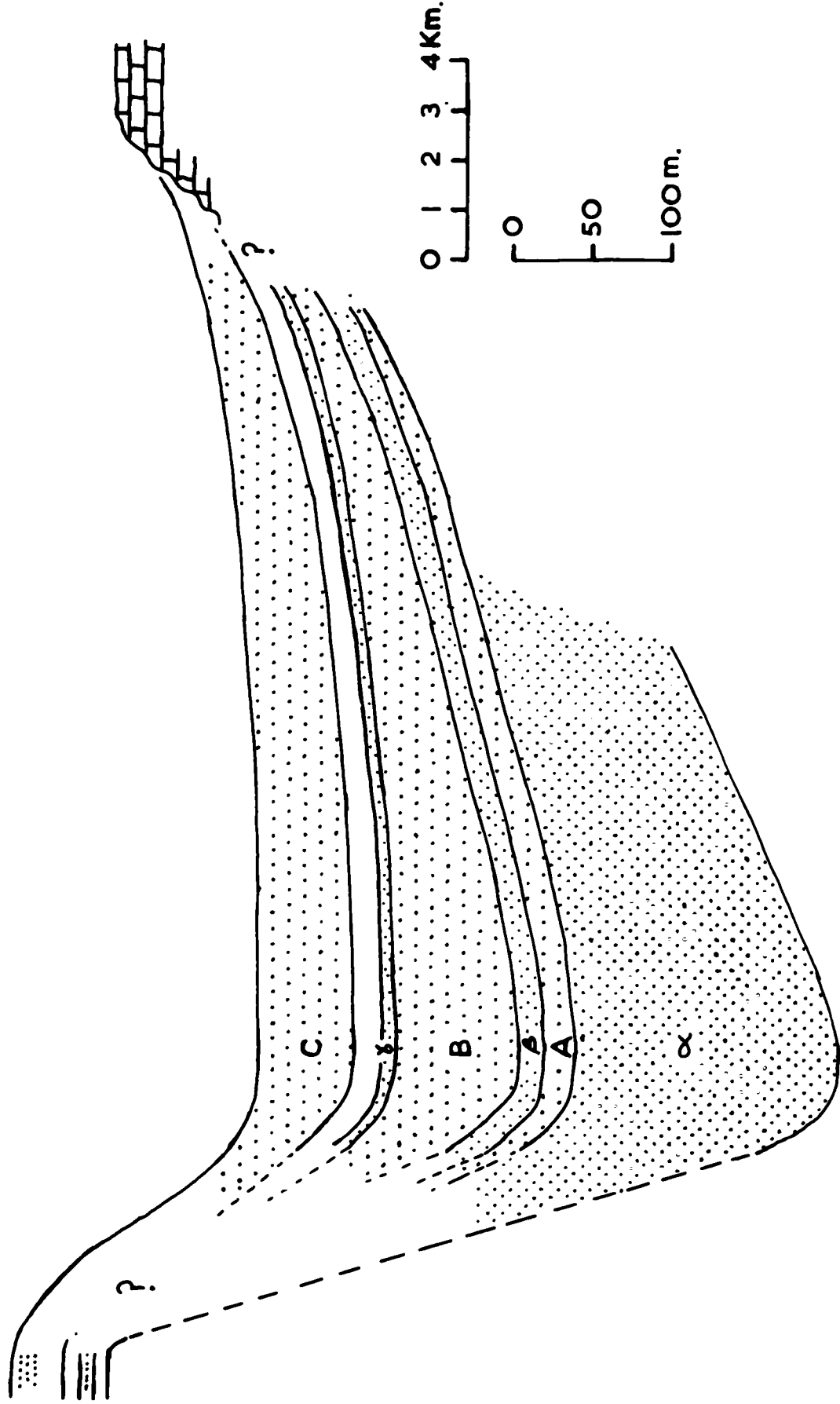
deposition at Astbury 4 miles to the west, whereas they are seen in an attenuated form 9 miles to the east at Elkstones. This indicates that apart from the greater subsidence taking place in the Lask Edge area there was in fact a physical slope of the sea bed up to Astbury which stopped the turbidity currents reaching the Astbury area. The slumping from the west observed in the Croker Hill area confirms the presence of a slope at the western margin of the basin area. Fig. 4b illustrates the general shape of the basin cross section during deposition of the E2 turbidites together with the relative thicknesses of the deposited sediments.

To the south of Lask Edge no exposures are available in E1 and E2 and the manner in which the lower Namurian sediments die out towards the land mass which lay to the south is not known. In north Derbyshire Walker (1966) has recognised channels in the Shale Grit which he believes acted as feeder channels for the turbidites of Mam Tor. He also recognises in the proximal turbidite developments deposits similar to those recorded by Gorsline and Emery (1959), Shepard and Einsele (1962), and Hand and Emery (1964) from the submarine fans at the bases of canyons off the California coast. While there are certain similarities between the proximal protoquartzites and such submarine fan deposits, too little is known about the areal variation

WEST - EAST CROSS SECTION OF NORTH STAFFORDSHIRE BASIN

Fig. 4b

ASTBURY	LASK EDGE	HURDLOW	BLAKE	MASSIF EDGE
W.				E.
WATER SURFACE				



of the beds in Staffordshire to permit such a close comparison. It must be concluded that the proximal protoquartzites were formed near the base of the slope on which they were generated but it is not known what form this slope took. It is probable that the slope was connected with the southern margin of the basin area but borehole evidence to the south does not give any indication of where this might have been situated.

The only borehole which provides any information was sunk at Werrington (Gifford 1923 p.239) and is described as starting below the Third Grit and penetrating 810 metres of 'Pendleside' and 'Yoredale' rocks. It would seem therefore to be in basin facies as Carboniferous limestone was not reached. A borehole at Chartley (about 6 miles NE of Stafford) contained 'Pendleside' facies resting on Carboniferous limestone which contained marls sandstones and conglomerates indicating probable proximity to a shoreline in the upper Viséan. Farther to the south in the Cannock and Wolverhampton areas Coal Measures rest directly on pre-Carboniferous rocks.

It would be particularly interesting to know what changes take place in the upper Viséan and lower Namurian to the south of Stoke-on-Trent. This area conceals the southern margin of the basin and the source area of the turbidites. Whether or not slope deposits and delta

deposits of E2 age would be found is a matter for conjecture, since the lower Namurian and upper Viséan are usually the first strata to be cut out by unconformity at the basin margins.

Area to the West of the North Staffordshire Basin.

The Astbury section has already been described as situated on the western margin of the basin, and boreholes to the south-west of Astbury confirm this view. At Bowsey Wood (Wrine Hill) Carboniferous limestone was encountered close below the G. cancellatum band with an unconformity between them (Earp and Calver 1961 p.182). Only four miles to the east of Bowsey Wood the Apedale borehole (Gifford 1923 p. 239) entered tuff just below the base of the Coal Measures and continued in tuff for 845 metres. This volcanic material was presumed to be Carboniferous in age, and if this is the case there is a remarkable change from here to Bowsey Wood.

Between Market Drayton and Newport three boreholes have been sunk (Wills 1956, Appendix A). No definite Namurian occurs in any of them and one at Ternhill has Upper Coal Measures resting unconformable on Lower Carboniferous. At Egmond near Newport, Coal Measures rest on pre-Cambrian, the Lower Carboniferous being unrepresented.

It is not known what happens to the Namurian

beneath the Cheshire Plain where the New Red Sandstone is very thick, but west of this area in Denbighshire and Flintshire and also in the Formby borehole an unusual Namurian succession is seen. Eumorphoceras bisulcatum has been recorded from the Holywell Shales of Flintshire by Jones and Lloyd (1943). Their locality for E. bisulcatum in the Terrig River has been visited by the author and the horizon is of E2b2 age as it contains E. aff. leitricense and Ct. aff. nitidus. Wood (1936) records E. bisulcatum associated with P. cf. membranacea and crinoid ossicles from the same area. It is possible that this represents the main E2a2 faunal band which often contains crinoids. To the south E. bisulcatum has been recorded from within the Cefn-y-Ffedw Sandstone (Wood 1936 p.16) and Wood considered that the sandstone was the lateral equivalent of the Holywell Shales to the north.

The Cefn-y-Ffedw Sandstone contains some shelly marine sandstones and was deposited in close proximity to a shoreline. Jones and Lloyd (1943) considered that there was an unconformity present beneath the "E zone" in Flintshire and this appears to be so. It is also certain that the Namurian in Flintshire overlaps the Lower Carboniferous since G. cancellatum has been found close to the Lower Carboniferous in Nant-y-Figillt, NW of Mold (Wood 1936 p.20). To the north of these

exposures - which show the overlap of the Namurian on the Lower Carboniferous of Wales, and the development of marine shelly sandstones in the Lower Namurian - a borehole at Formby described by Kent (1948) provided an unusual succession. In the borehole c.381 metres of beds were ascribed to the upper Carboniferous since they occur above beds with P2 goniatites. The beds were calcareous or dolomitic throughout and contained calcareous sandstones and siltstones with shales and limestones. A shelly marine fauna occurred at some levels and Kent concluded that at least half and possibly all the succession was marine. Kent provisionally ascribed the succession to E zone and correlated it with the Cefn-y-Ffedw Sandstone. The suggested correlation by Hudson (in Kent 1948 p.262) of these beds with the Gun Hill succession has no palaeontological basis, but it is interesting to note that the derived fauna of crinoids, bryozoa, Endothyrid foraminifera and sponge spicules found in the Gun Hill Siltstones Formation are all recorded from these beds in the Formby borehole. It was considered earlier that the Staffordshire calcareous siltstones were derived from an area where shallow water shelly sandstones were accumulating, and this succession seems to be of a suitable type to provide the material for the calcareous siltstones of north Staffordshire. The thick nature of the Formby succession is continued into P2 but it seems

that sedimentation roughly kept pace with subsidence to maintain shallow water conditions with a shelly benthonic fauna in the upper Viséan and lower Namurian. Possibly some of this material was swept eastwards into the north Staffordshire basin to form the calcareous siltstones of the Gun Hill Siltstones Formation, and into north Derbyshire to form the Alport Crowstones of Hudson and Cotton (1943).

Area to the North of the Staffordshire Basin.

The protoquartzitic turbidites of the Staffordshire basin are not seen again north of the Croker Hill area. They are represented in Edale by shale-mudstone with siderite and no trace of turbidite material is seen. It is evident that the turbidity currents which were flowing northward at Croker Hill did not flow into the north Derbyshire area and this was possibly due to a submarine barrier between the two areas over which the currents could not flow.

The situation at the margin of the limestone massif in north Derbyshire is similar to that seen in Staffordshire with overstep of the Namurian on to the lower Carboniferous and thickening of the sediments away from the massif into the basin area (Kent 1966). Kent (1966) recognises two lobes of thick Namurian sediments extending ESE into the northern edge of the massif area, these lobes he refers to as the Edale Gulf

and Gainsborough Trough, and they were probably being filled with sediment during E1 and E2 times.

The Derbyshire Limestone Massif Area.

During Viséan times the edge of the massif was marked by reef limestones such as those seen at Treak Cliff, Castleton; at Parkhouse Hill and Chrome Hill in the Upper Dove; Narrowdale Hill near Wetton, and in Dovedale. On the massif area pale, well bedded limestones are the usual lithofacies whilst in the basin areas thin bedded dark limestones and shales accumulated. Prior to the deposition of the Namurian, limestone deposition ceased and some erosion of the reef margins took place. The Namurian shales were then deposited unconformably on the margins of the massif area so that there is an overlap of the Namurian on to the massif.

This is well seen in E zone in the Upper Dove and at Mam Tor, Derbyshire,

On the top of the Derbyshire massif area the lower Namurian sedimentary cover is very thin where preserved and may be absent due to unconformity. The Ashover boreholes (Ramsbottom et al. 1962) contained extremely thin developments of E zone with development of phosphatic nodules in E1. The sediments are water laid, but very little detritus reached this area during the lower Namurian. Thicknesses of the successions from C. leion

to Ct. edalensis range from 10 metres to 12.5 metres in the three boreholes. A similar thin succession is present at Wardlow Mires (Inst. Geol. Sciences Annual Report 1966 p.71) where only 4.6 metres separates E. bisulcatum from E. pseudobilingue. Further to the east at Eakring which lies at the head of the Edale Gulf no dateable horizons are present between R1 and the Viséan. In view of the extremely thin nature of the E zone sediments at Ashover it is likely that they would be easily eroded even if deposited at all over the massif area, and an unconformity or non-sequence may be present. Localities to the south of Newark-on-Trent (Falcon and Kent 1960) again show thin developments of E and P sediments or complete absence of these zones.

The Windmerpool Gulf.

This basin or gulf area is situated along the southern margin of the massif area and contains a typical thick development of P and E sediments (Falcon and Kent 1960, Kent 1967) which have been proved in borings at Windmerpool, Long Clawson and Hathern. Recently a borehole at Duffield has proved a similar thick succession with units of graded sandstones and siltstones, probably turbidites, present in E2. The Ct. edalensis horizon is separated from a horizon with E. bisulcatum and C. cowlingsense by 53 metres of mudstones

with graded sandstone and siltstone, and a further thick unit of graded beds is present beneath the E. bisulcatum horizon (Inst. Geol. Science Annual Report 1966 p.71). It is not yet known if these units are exactly synchronous with those in north Staffordshire or whether they come from the same general source area. The probable extent of the Windmerpool Gulf based on the surface topography of the Carboniferous rocks (Kent 1967) is shown in Fig.4a. Kent (1967 p. 130) mentions the problem of the connection between the Windmerpool Gulf and the north Staffordshire Gulf. Evidence from the Duffield borehole may help to solve the problem but until deep boreholes are put down in the region of the supposed connection the question will remain in doubt. In view of the fact that the greatest thickness of E sediments in north Staffordshire occurs along a N-S line at the west of the basin makes it seem likely that this was the main axis of the Staffordshire basin, and it is unlikely that this connects directly with the E-W trending Windmerpool Gulf. A submarine barrier possibly separated the two as shown in Fig. 4a. To the south of Windmerpool Gulf rapid thinning of the Namurian takes place, especially in the region of Charnwood Forest. E and P zones are usually eliminated by unconformity and upper Namurian rests on thin Lower Carboniferous or pre-Cambrian.

Conclusion.

The north Staffordshire basin area is similar to the Windmerpool and Edale Gulfs in containing thick developments of E and P sediments of basin facies. All three basin areas are developed marginally to the Derbyshire massif area which received little sediment during E and P times and remained relatively stable while the surrounding basins subsided to accommodate a large thickness of sediment. Rapid attenuation of the lower Namurian occurs to the south of the Windmerpool Gulf and to the west of the Staffordshire basin usually with elimination of P and E deposits by unconformity. Further to the west in Denbighshire and Flintshire lower Namurian is again developed and can be demonstrated to overlap the Carboniferous with unconformity. Shelly sands developed here and those seen in a borehole at Formby (Kent 1948) provide a possible source material for the calcareous siltstones of the north Staffordshire basin area. The deltaic deposits that supplied the material for the protoquartzitic turbidites in the north Staffordshire basin are possibly concealed under the Coal Measures south of Stoke-on-Trent but may have been removed by erosion since this part of the succession is frequently cut out by unconformity at the basin margins.

At the same time as basins were being developed and filled with sediment around the Derbyshire massif,

and unconformities forming in areas marginal to the basins or on the massif areas, significant changes also took place further north. On the Askrigg Block an intra-E1 unconformity provides evidence of movement during E1 time which locally may be responsible for the unconformable relationship of upper E1 upon Viséan D1 limestones.

During upper Viséan and lower Namurian times earth movements took place which affected sedimentation on the relatively stable massif areas. Local uplift produced unconformities and in areas of little clastic supply stable conditions gave thin complete shale sequences. Marginal to the massif areas rapid subsidence and sedimentation took place, unconformities were developed at the massif margins and rapid changes in sediment type and thickness occurred across the margins. In some cases faulting of the basin margin can be demonstrated (e.g. Craven Faults) to have taken place but in others, (e.g. north Derbyshire, north Staffordshire) a simple monoclinial flexuring of the basin margin has taken place, whether or not this is connected with faulting in depth is not known.

Superimposed on this tectonic activity were further changes affecting a very large area and producing a type of cyclic sedimentation in the north Staffordshire basin area which is described in the following chapter.

PART 5.CYCLIC SEDIMENTATION AND ENVIRONMENTAL INTERPRETATIONS.

- 5.1 Descriptions of cycles observed.
- 5.2 Comparison of the north Staffordshire basin cycles with some relevant contemporaneous forms of Carboniferous cyclic sedimentation.
- 5.3 Environmental interpretations.
- 5.4 Possible causes of the cyclic sedimentation observed.

CYCLIC SEDIMENTATION AND ENVIRONMENTAL INTERPRETATIONS.

5.1

DESCRIPTION OF CYCLES OBSERVED.

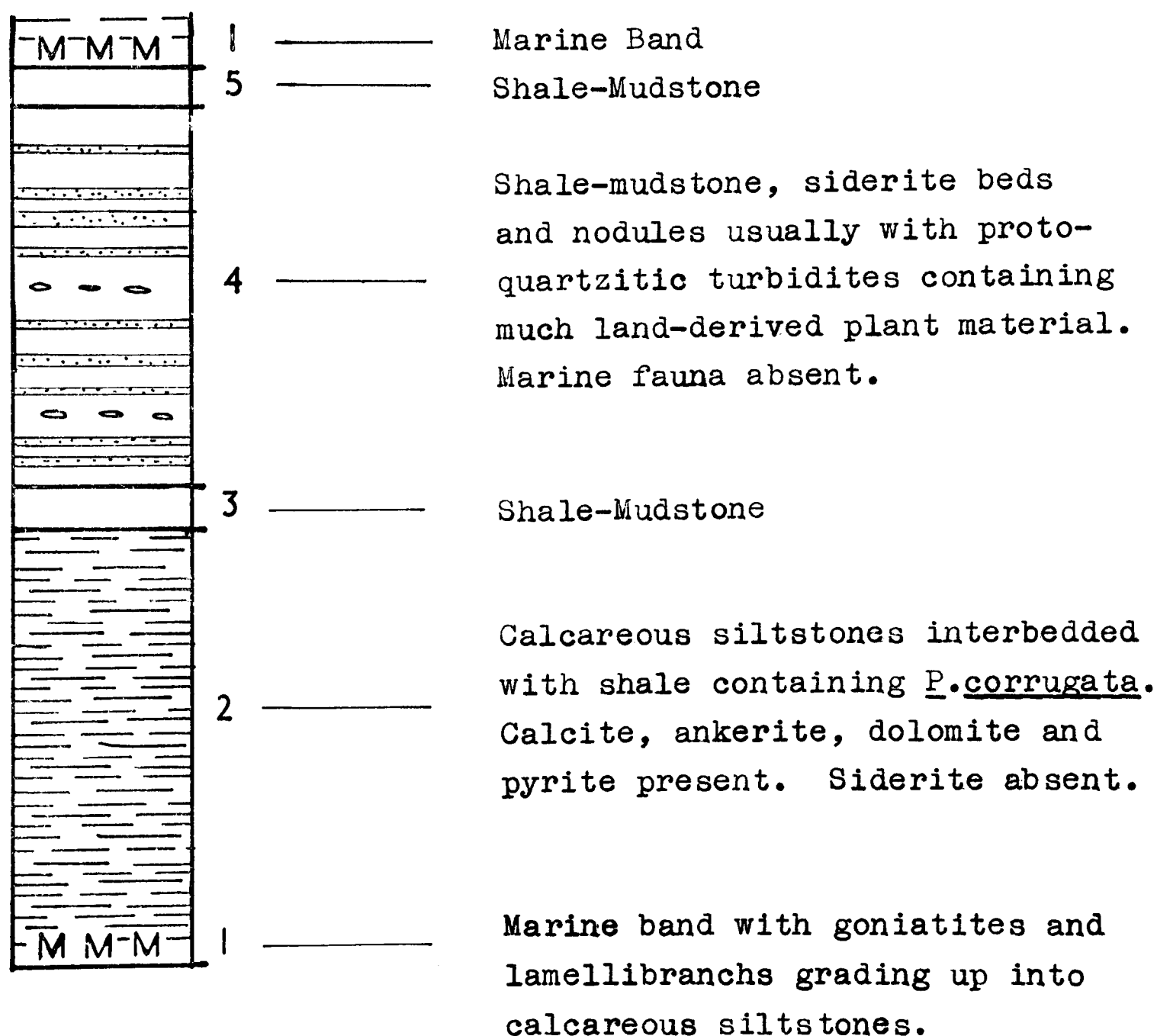
Within the part of the succession studied cyclic sedimentation has been recognised by the author. The typical form of the cycle seen is shown diagrammatically in Fig.5.1a.

Within this cycle there are no major erosion surfaces or reworked horizons other than those immediately below individual turbidites, and there is no evidence that emergence ever took place during deposition of the cycle. The sedimentology of both the protoquartzites and the calcareous siltstones already described, indicate deposition in a body of water of such a depth that wave action was absent. The water was sufficiently deep during deposition of the upper, turbidite bearing part of the cycle for turbidity currents to develop in a shallower area and flow downslope into the basin.

At the margin of the basin in the Upper Dove, at Stannery, the Viséan reef remnants of Chrome Hill and Parkhouse Hill have a relief of nearly 120 metres above the E2a Namurian resting unconformably on their flanks. The fact that thin Namurian deposits occur on the massif area, as at Wardlow Mires and Ashover, indicates that it was covered with water, and thus the depth of the

Fig.5.1a.

Depositional cycle. N. Staffs. Basin.



water at the margin of the basin must have been at least 120 metres, assuming that little subsequent tectonic activity has affected the margin of the massif area. The fact that there is no evidence of contemporaneous E2 wave erosion at the margin of the basin or on the massif area nearby may mean the water depth was considerably greater. To the east, on the massif area at Eakring, no dateable faunas occur between the Viséan and R1 subzone, and further east near Lincoln unconformity does occur eliminating E zone, and thus locally emergence may have taken place to the east. It seems likely that the depth of water over the massif area was variable but not very great, possibly up to 200 metres. This gives a possible depth of up to 320 metres of water near the basin margin in the Upper Dove.

In the succession studied, together with the Gun Hill Siltstones Formation, there are four apparent cycles conforming to the general plan of Fig.5.1a. The proportions of the siderite and calcite bearing parts of the cycles vary considerably (Fig.5.1b). There is a marked reduction in thickness of the calcitic parts of the cycles from the base upwards, and with the exception of the turbidite free, shale-mudstone with siderite unit beneath the E2a2 faunal band, there is an upward thickening of the siderite bearing parts of the cycles. There is thus a general change from predominant calcareous siltstone

to predominant protoquartzite deposition from E1 to E2b.

It is difficult to estimate the relative proportions of time involved in the deposition of the various parts of the cycles. The time represented by the protoquartzitic turbidite beds is negligible and it is probable that the time represented by the individual calcareous siltstone beds is also negligible geologically, although they may have taken longer to be deposited than the protoquartzitic turbidites. The latter were probably deposited in a matter of hours (Kuenen 1967) whereas the calcareous siltstones may have taken a few days.

If the total thicknesses of the turbidites and the calcareous siltstone beds in the various units are roughly calculated, (by using measured logs of parts of the units) and subtracted from the thickness of the unit, the thickness of pelagic shale and shale-mudstone present is obtained. The thicknesses so obtained for the Pyeclogh section are shown in Fig.5.1c, from which it can be seen that the shale-mudstone thicknesses of the siderite bearing parts of the units increase progressively up the succession and those of the calcitic parts decrease in the same direction.

It is not possible to determine the relative rates of deposition of the shales and shale-mudstones found in the marine bands, between the calcareous siltstones, and between the protoquartzitic turbidites associated

with siderite, but the following considerations lead to the conclusion that the sideritic parts of the units were deposited more quickly than the calcitic.

The state of the clay minerals in the three types indicates that those in the marine shales have been most altered and were probably nearer stability with the marine environment than those associated with the protoquartzites. This probably indicates slower deposition for the marine band shales than for the shale-mudstone with siderite, where the original detrital clay fraction of kaolinite, illite and some chlorite typical of the coal measures is preserved.

Taylor and Spears (1967) and Curtis (1967) consider that the formation of diagenetic pyrite or siderite in a sediment is controlled by the concentration of dissolved sulphur species. If the rate of sedimentation was quick, the concentration of sulphur species produced by sulphate reducing bacteria would be low and siderite would tend to form. If the concentration of sulphur species was high, as in slower deposition, pyrite would form.

This idea correlates well with the present section where the sideritic part of the succession shows evidence on the basis of clay mineralogy of faster deposition than the lower calcitic (and pyritic) parts of the cycles. The association of siderite with clearly land derived deltaic material is similar to that noted for the

Yorkshire Lias by Hallam (1967) where calcite is similarly associated with transgressive phases of the cycles.

Curtis (1967 p.2122) considers that both pyrite and siderite formed in a diagenetic physiochemical environment of pH 7-8 and Eh -0.2 to -0.3V and these conditions spanned marine, brackish and non-marine environments as determined by the faunal changes. Nicholls and Loring (1962) considered changes in pH and Eh caused change from pyrite to siderite but found no geochemical evidence of overall change in salinity over a coal measures cycle. There is thus no real evidence that there were any marked salinity changes associated with the coal measures cycles and the parts of cycles referred to as 'marine' or 'non-marine' are taken to refer to the presence of a marine fauna, and in the case of non-marine to the presence of 'non-marine lamellibranchs' or the absence of a marine fauna.

If it is taken that the shale in the siderite bearing part of the cycle accumulated more rapidly, it must be remembered that the rate of pelagic shale deposition - sideritic or calcitic - over the area was not constant for given periods of time, as is shown by the distances of separation of K-bentonites B4, B5 and B6 (see Part 1.4) at various localities. However it does seem, from the same evidence, that at any one place the

rate of shale deposition was fairly constant, and thus thicknesses of pelagic shale could be related to time in a single section, but comparison with other sections would not be possible without correcting for different rates of accumulation. In the Pyecclough section, it can be seen from Fig.5.1c that on this basis there is a progressive change in the four cycles recognised in the proportions of time devoted to the calcitic and sideritic parts of the cycles.

In the north Staffordshire basin area no goniatite bearing horizons are known between the C.leion and C. malhamense horizons. Several bands would be expected in this interval containing E. pseudobilingue faunas (cf. Yates 1962). In Edale, such an horizon is found at Loc. 78 of Hudson and Cotton (1945). The goniatites occur in two rows of bullions in the base of a unit of calcareous siltstones eight metres thick overlain by shale-mudstone which is the equivalent of Unit A proto-quartzites in Staffordshire. The amount of shale in the calcareous siltstone unit above the E.pseudobilingue band was calculated and the thickness corrected for the different rates of deposition in Edale and at Pyecclough by comparison of Unit β at each locality. The shale equivalent thicknesses then read as in Fig.5.1d.

Fig.5.1b

Total thicknesses of calcitic and sideritic parts of cycles at Pyeclough.

	Calcitic. cms.	Sideritic + junction shale. cms.
<u>E. erinense</u> at base	488	3812
<u>E. grassingtonense</u>	915	1220
<u>C. malhamense</u>	1,464	2074
<u>C. leion</u> at base	15,250	1220

Fig.5.1c

Thicknesses of pelagic shales in sideritic and calcitic parts of the cycles in Pyeclough section.

	Calcitic shale. cms.	Sideritic + junction shale. cms.
<u>E. erinense</u> at base	122	1525
<u>E. grassingtonense</u> at base	229	1220
<u>C. malhamense</u> at base	366	824
<u>C. leion</u> at base	3812	488

The marine bands and calcareous siltstone units were treated in the same way and were estimated to consist of 25% marine pelagic shale. The protoquartzitic turbidite bearing units were considered to consist of 40% black basinal shale- mudstone (pelagic origin).

Fig.5.1d

Relative thicknesses of pelagic shale in the calcareous and non-calcareous parts of the cycles (based on Pyeclough and Edale sections).

	Calcareous part of cycle.	Non-Calcareous (mainly siderite bearing).
<u>E. erinense</u> cycle	122	1525
<u>E. grassingtonense</u> cycle	229	1220
<u>C. malhamense</u> cycle	366	824
<u>E. pseudobilingue</u> cycle (Edale only)	488	488
Gun Hill Siltstones	3324	-
<u>C. leion</u> at base		

It has already been suggested that the rate of deposition of marine shale was less than that of the shale in the sideritic units on the basis of the clay mineralogy and diagenetic mineralogy. In order to find the rate of calcareous shale deposition relative to that of the black shale-mudstone an assumption must be made concerning the relative lengths of time occupied by the cycles. If it is assumed that the four complete cycles in Fig.5.1d each accumulated over a equal period of time, and that pelagic shale deposition was directly proportional to time, then an attempt can be made to relate the rate of marine (calcareous) shale deposition to that of the black shale-mudstone in the siderite bearing units. There is no real evidence to back this assumption, but the resulting

relative rates of deposition of the two types of pelagic sediment are then similar for each cycle, and so the assumption, large though it is, may be valid. When the amount of calcareous shale is multiplied by a factor of from 2.3-3.3 and added to the amount of non-calcareous shale, then the four resulting figures for the cycles are reasonably constant. The 'best' factor is 2.85 which gives the figures in Fig.5.1e for the 'time-equivalents' of each cycle.

Fig.5.1e

Calc. shale thickness	Calc. shale x 2.85	Sideritic shale thickness	'Time units' (Calc. shale x 2.85 + sideritic shale)
122	348	1525	1873
229	653	1220	1873
366	1043	824	1867
488	1391	488	1879

On this basis it seems possible that if the four cycles did occupy roughly equal time intervals, then the calcareous shale was deposited at about $\frac{1}{3}$ the rate of the shale-mudstone in the sideritic units. It is interesting to note that the shale equivalent of 3324 (Fig.5.1d) for the Gun Hill Siltstones would seem to represent a much larger span of time than that of the four complete cycles. Using the figures of Fig.5.1e, the number of 'time units' represented in the Gun Hill Siltstones is $3324 \times 2.85 = 9,475$

which when divided by 1873 (the average number of 'time units' in a cycle from Fig.5.1e) gives 5.06, a remarkable approach to a whole number multiple of the cycles recognised. If the cycles did each last the same length of time, then it is suggested that five 'cycles' are present in the Gun Hill Siltstones Formation, but that conditions were never suitable for siderite deposition, as there was never any influx of deltaic material into the basin. The marine bands which would be expected to occur have not been found in the basin area but a number of faunal bands do occur in other areas (e.g. Leitrim, Yates 1962) between C. leion and C. malhamense and contain varieties of E. pseudobilingue, E. medusa, E. angustum, etc.

This treatment of the cycles by dividing them into calcareous and non-calcareous parts and concerning primarily one section (Pyeclogh) gives simple results. Two factors which may contribute to reduce the significance of these preliminary treatments are as follows.

(1) It has been assumed in calculations that the overall rate of deposition of marine bands was the same as that of the calcareous siltstones. This is probably not the case, the marine bands being deposited more slowly since they contain a greater proportion of pelagic shale than the calcareous siltstone units. A more sophisticated calculation could treat the goniatite bearing marine bands and the calcareous siltstones differently.

(2) Between the calcareous siltstone and the proto-quartzitic turbidite parts of a cycle there is always about a metre of shale which contains no coarse grained beds and lacks both sideritic and calcareous concretions. Somewhere within this interval the change over from conditions favouring calcite - ankerite - dolomite to those favouring siderite deposition is recorded. In the calculations this interval has been treated together with the sideritic parts of the succession. A similar shale transition occurs from the top of a siderite bearing unit to a marine band. These thicknesses of shale are important in marking changing conditions and probably took considerable periods of time to be deposited.

In view of these two factors a slightly more elaborate calculation of relative time of accumulation of the cycles can be attempted. In Fig.5.1f, the thicknesses of the units at Pyeclough are shown with the marine bands and shale transitions shown separately. In Fig.5.1g, the shale equivalents are calculated on the same basis as before with the exception that the marine bands are considered to consist of 50% pelagic shale (instead of 25%). The 'time equivalents' calculated (Fig.5.1h) are still very similar and tend to indicate that a similar period of time was involved in the deposition of each cycle.

When the time equivalents of the various cycles are

Fig.5.1f

Actual thicknesses(cms) of parts of units.

	Marine Band.	Calcareous Siltstones.	Shale Transition.	Siderite bearing.	Shale Transition.
<u>E. erinense</u>	300	-	100	3750	180
<u>E. grassingtonense</u>	130	900	← 1200 →	1200	→
<u>C. malhamense</u>	50	1450	60	2000	260
<u>E. pseudobilingue</u>	40*	1920*	140	1200	120

* estimated from Edale succession and corrected for different thickness in Edale and Pyeclough.

Fig.5.1g

Pelagic shale equivalent thickness of parts of units (cms.).

	Marine Band.	Calcareous Siltstones.	Shale Transition.	Siderite bearing.	Shale Transition.
<u>E. erinense</u>	150	-	100	1500	180
<u>E. grassingtonense</u>	65	225	← 1200 →	1200	→
<u>C. malhamense</u>	25	362	60	800	260
<u>E. pseudobilingue</u>	20	480	140	480	120

Fig.5.1h

Time equivalents of pelagic shale deposition. Marine shale x 3.

	Marine Band.	Calcareous Shale	Siderite. Shale	Total.
	Siltstones.	Transition.	Transition.	'Time Units'
<u>E. erinense</u>	450	100	1500	= 2230
<u>E. grassingtonense</u>	195	675	1200	= 2070
<u>C. malhamense</u>	75	1086	800	= 2281
<u>E. pseudobilingue</u>	60	1440	480	= 2240

Fig.5.1i

Time equivalent calculations for cycles at various localities.

Cycle	<u>E. pseudobilingue.</u>	<u>C. malhamense.</u>	<u>E. grassingtonense.</u>	<u>E. erinense.</u>
Loc				
Blake	2240	2345	1770	2160
Sunnydale			1530	
Hurdlow			1560	
Elkstones		2225	1405	
Pyeclogh	2240	2281	2070	2230
Upper Dove				1425
Edale	1260	1280	1700	1950

If all the units are of approximately the same duration then the low figures obtained indicate slower deposition at a particular locality.

calculated on this basis for other localities in north Staffordshire, similar figures are obtained which tend to confirm the suggestion that the cycles each took approximately the same amount of time to be deposited (Fig.5.1i).

It is of importance to note that when shale equivalents are calculated, the shale transitions become relatively more important and constitute 10-20% of a cycle (Fig.5.1g). The change over from one type of deposition to another was not sudden but took place very slowly with a considerable time gap between the cessation of calcareous siltstone deposition and the start of proto-quartzitic turbidite deposition associated with siderite.

Minor Cycles.

Within both the calcareous siltstone and the protoquartzitic turbidite units deposition of coarse beds appears to follow a cyclic pattern. The thicker beds of both lithofacies tend to be grouped together within the unit and the groups are separated by shale or shale-mudstone with thin beds of the coarser lithology. The cumulative diagrams of turbidite/black shale-mudstone and calcareous siltstone/shale (Figs.2.1Ca, 2.4Ha) illustrate this point.

It has not been possible to determine the exact number of minor cycles within a major cycle due to poor exposure, and it is not known whether there is a constant number of minor cycles in a major cycle. In the

calcareous siltstones Unit β at Oakenclough (151), each minor cycle was associated with about 60cm. of shale deposition. On this basis there should be about six such cycles in Unit β of the calcareous siltstones. In Unit B protoquartzites at Fox Bank Quarry, five minor cycles can be seen in the quarry face in what must be a nearly complete section of Unit B. It thus appears that within the major cycle with C. malhamense at the base there are at least 11 minor cycles present and others are likely to be masked in the shale intervals between the units bearing the coarser grained lithofacies. It is thus tentatively suggested that at least 11 minor cycles may constitute the major cycle. Detailed logging of complete borehole sections could provide interesting information on the relation of major and minor cycles, but no such cored sections have been available to the author. The borehole at Duffield, Derbyshire (34284217) (Harrison in discussion of Trewin 1968) may provide information of this kind.

Summary.

In the succession studied there is an overall trend from calcareous siltstone deposition at the base to protoquartzitic turbidite deposition at the top. This is part of the general trend of sedimentation in the Staffordshire basin area which starts with basinal marine limestones and shales in the Viséan and ends with typical

coal measures deposition in the Westphalian. Superimposed on this general trend is the cyclic sedimentation described which produces the typical widespread marine bands of the Carboniferous. The minor cycles recognised appear to affect the deposition of the coarser clastics but are of minor importance in their effect on the major cycles. It is possible that the major cycles were all of approximately the same duration and that the rate of pelagic marine shale deposition was $\frac{1}{3}$ that of the shale-mudstone with siderite nodules. The division of the cycles into a lower calcite-ankerite-pyrite bearing part and an upper siderite bearing part is attributed to a lower concentration of sulphur species produced by sulphate reducing bacteria in the faster deposited sideritic part of the cycle.

5.2

COMPARISON OF THE NORTH STAFFORDSHIRE BASIN CYCLES WITH
SOME RELEVANT CONTEMPORANEOUS FORMS OF CARBONIFEROUS
CYCLIC SEDIMENTATION.

In order to interpret the mode of deposition of the cycles seen in the north Staffordshire basin area, it is necessary to review other forms of cyclic sedimentation affecting clastic deposition in the Carboniferous, and to attempt to correlate such cycles with events in the north Staffordshire basin area. The best known cyclic sequences are those of 'Yoredale' and 'Coal Measures' types. It must be recognised that many other types of cycles have been described from the Carboniferous (Duff et al. 1967, Ch. 5 for summary). Basically the cycles described in the literature range from a marine horizon, through clastic deposition to seatearth and coal.

Sections of Yoredale and Coal Measures cycles are shown in Fig.5.2a and are compared with the Namurian basin cycle described previously. The marine lower portions of the cycles are very similar with the greatest abundance and variety of marine forms concentrated at the base of the cycle. Upwards the marine faunal element diminishes until there is predominant deposition of immediately land derived material before the next marine fauna appears. There is a marked difference between the sedimentology of the rocks in the north Staffordshire

Fig 5.2a

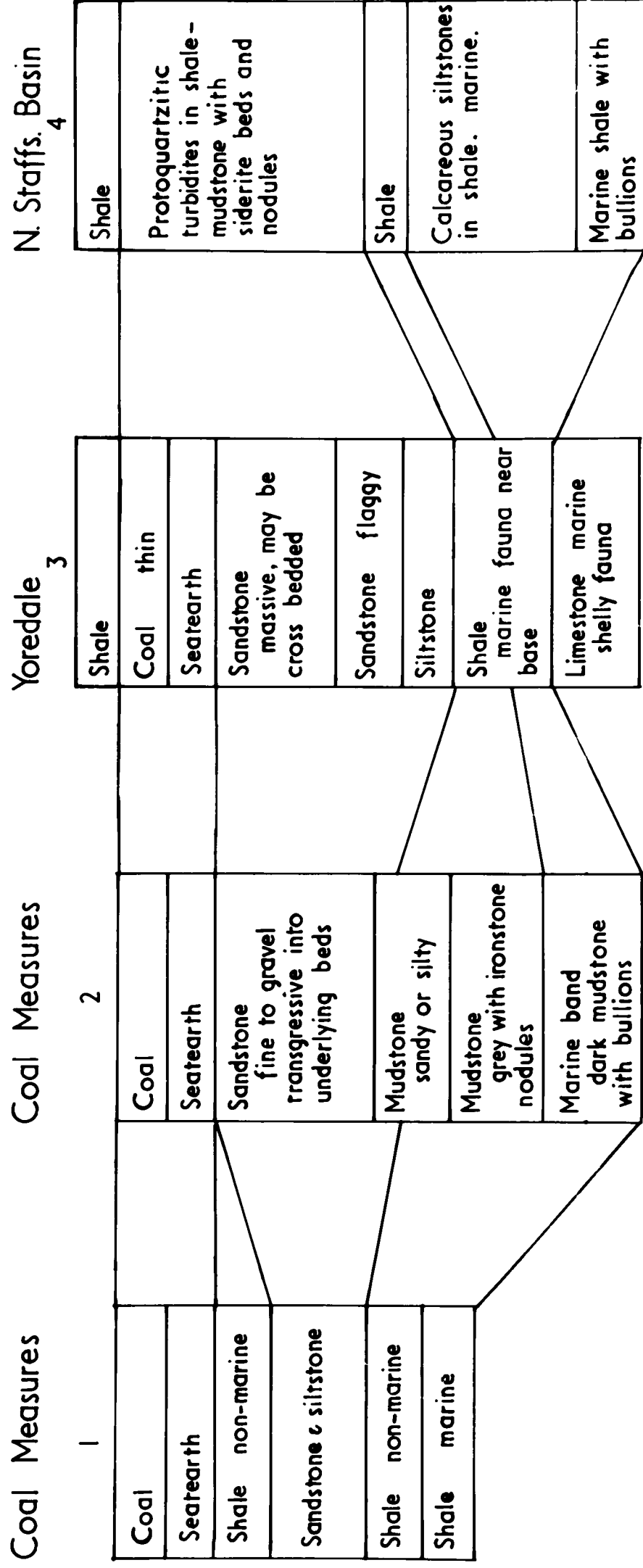


Diagram to show the equivalent lithologies of cycles of Coal Measures and Yoredale type compared with the north Staffordshire basin cycles.

basin environment and those of the shallow water 'Yoredale' or 'Coal Measures' cycles. The shallow water cycles are characterised by a rapid and widespread transgression covering a large area and establishing shallow water marine conditions (Wells 1960 p.391). Deposition of the clastics is generally assumed to be rapid and due to deltas building out from the shoreline and covering the marine sediments first with pro-delta clays, and then with the sands and silts of the advancing delta slope. The development of seatearth and coal took place slowly at about sea level on low lying swamps bordering the coast (Johnson 1960). On this scheme a long period of relative quiescence follows a rapid transgression of the sea which covers a large area previously occupied by coal producing swamp.

The composite cycles illustrated for the Yoredale and Coal Measures sedimentation are usually imperfectly developed, and may not actually occur in a given sequence. The most frequently developed cycle (Duff and Walton 1962 p.239), usually consists of only a part of the composite cycle. Duff and Walton (1962 p.244) found that the three most commonly occurring cycles in the Coal Measures of the East Pennine Coalfield are as follows:-

<u>Primary mode.</u>	<u>Secondary mode 1.</u>	<u>Secondary mode 2.</u>
Coal	Coal	Coal
Seatearth	Seatearth	Seatearth
Shale	Shale	Shale
	Siltstone	Siltstone + sst.
	Shale	Shale

In these cycles there is an emphasis on the non-marine conditions of seatearth and coal formation which are represented in all three modal cycles rather than the marine conditions which are not represented in any of the modal cycles of this region. In the Yoredale type of succession, the marine portion of the cycles is better, and more frequently developed and a statistical study of these beds may reveal that the modal cycle includes the marine conditions rather than the non-marine.

Before comparison is made between the shallow water cycles and the Namurian basin cycles it is necessary to find an area where Yoredale or Coal Measures type cycles were accumulating during the Namurian and in particular during E1 and E2 so that it can be shown that shallow water cycles were being produced at the same time as those in the basin area.

In the Yoredale cyclothem deposits in the north Pennines on the Alston-Askrigg block the marine faunas present at the bases of the cyclothems generally lack goniatites, thus direct correlation with goniatite bearing marine bands is usually impossible. The base of

the Namurian is taken at the base of the Great Limestone (Johnson et al. 1962 p.357) on the basis of the presence of C. leion above this limestone (Ibid p.344). C. aff. malhamense is recorded by the same authors from above the Little Limestone in the Swinhopehead area, East Allendale, but as mentioned in section 1.3 it is not identical with the C. malhamense of the present area. Above the Little Limestone, Dunham and Johnson (1962 p.250) record the presence of six cycles starting with shale or argillaceous limestone (marine or brackish) and proceeding through shale, shale + siltstone, sandstone and ganister to thin coal. These cycles are developed in the lower part of the Namurian but exact dating is not possible due to the lack of goniatites.

On the Askrigg block an intra-E1 unconformity is present which truncates the top of the Yoredale succession. The succeeding sediments of E1 and E2 age rest on all divisions of the Yoredales and in places on D1 limestones (Wilson 1960 Fig.7). In the Coverdale region (Wilson 1960) the Grassington Grit overlies the unconformity and is succeeded by the Cockhill Marine Band (E2a1).

Sections in the E2a and E2b strata of the Rombolds Moor area (Stephens et al. 1953) show marked cyclic development with the Edge Marine Band (E2a2) overlying a grit with a coal seam (the Bradley Coal). The marine band is overlain by shales and sandstones which pass

upwards into the massive Marchup Grit, and this grit is followed by marine bands which include the Ct. edalensis band. Wilson (1960a) shows a similar succession to exist in the Colsterdale region where the Cockhill Marine Band, correlated with E2a1, by Yates (1962 p.417) is succeeded by the predominantly shaly Nidderdale shales and then by the Red Scar Grit which is divided into two leaves, the lower being a coarse, massive, current bedded sandstone 4-53 feet thick (Wilson 1960a p.432) on top of which is a ganister and the Woogill Coal. Above the coal a fauna with Lingula and nuculids occurs in shale which is succeeded by the upper leaf of the Red Scar Grit which in places contains foreset beds up to 20 feet in amplitude (Ibid p.432). The Red Scar Grit is succeeded by marine beds of E2b1 age (Colsterdale Marine Beds).

It can therefore be briefly shown that cyclic sedimentation was taking place in the Rombolds Moor and Colsterdale areas during E2 and examination of other reports of successions in this part of England also show cyclic sedimentation to be developed. The succession in Ireland described by Yates (1962) lacks coarse grained deposition except in the interval from E. bisulcatum erinense (high E2a) to Ct. edalensis (E2b1).

The succession here is as follows:-

	Feet	Metres
<u>Ct. edalensis</u> E2b1	10	3.05
Shale	20	6.10
Massive grit	50	15.25
Rusty arenaceous shale + ironstone nodules	25	7.80
Coal	1	.30
Sandy shale + flagstones	50	15.25
Coal	3	.90
Massive grit + flagstones	75	23.05
Shale + clay ironstone bands	100	30.50
<u>E. bis. erinense</u> High E2a	10	3.05

Thus in Eire a similar cycle is developed to that in north England between the E2a2 and E2b1 faunal bands.

Around the margins of the Staffordshire basin area no exposures are available which show the near shore deposits of the lower Namurian. The nearest development is in H zone at Astbury on Congleton Edge where a cycle has been recognised culminating in ganister and coal deposition which is so rapidly followed by marine conditions that the topmost ganister bed is reworked and contains orthid brachiopods on its upper surface.

In shallow water areas where coarse clastic material was being introduced during E1 and E2 times there is good evidence that cyclic sedimentation of Coal Measures or Yoredale type was taking place, depending on the predominance of the marine horizons or the coals. It is

of interest to note that it is in E1 that the change occurs from predominant Yoredale type to Coal Measures type in the north of England and that it is at this level that the change from continuous marine deposition to marine with introduced land derived material occurs in the Staffordshire basin area.

A cyclic faunal development in a pure shale sequence at Ashover on the Derbyshire Massif has been noted by Ramsbottom et al. (1962) in which the following six 'faunal phases' were recognised, and attributed to increasing salinity although it was admitted that water depth may have been another important factor.

1. Fish phase
2. Planolites phase
3. Lingula phase
4. mollusc spat phase
5. Anthracoceras or Dimorphoceras phase (thin shelled goniatites)
6. Thick shelled goniatite phase

No definite cycles involving the above faunal phases can be correlated with the cycles recognised here since the E1-E2a part of the succession at Ashover is partly phosphatic and does not show a regular development of faunal phases. It can only be concluded that even in a pure shale sequence, variations, possibly cyclic in nature, in faunal development took place. These

variations could have been caused by a variety of factors such as salinity or other chemical variations in the overlying water. Equally the detrital material composing the shale may have been variable in composition, and affected the fauna. Only a detailed chemical investigation of the sediments could help to solve these points.

5.3

ENVIRONMENTAL INTERPRETATIONS.

If it is assumed that the shallow water equivalents of the basin cycles observed in north Staffordshire are of Coal Measures type, then an attempt can be made to explain the mode of introduction of the coarser material to the basin and the development of cycles in the basin.

It has already been shown that in the succession studied there is an overall trend in the sedimentation from calcareous siltstone deposition to protoquartzite deposition. Superimposed on the major trend is a series of changes giving cyclic sedimentation. The cyclic sedimentation will first be considered in relation to traditional 'Coal Measures' cycles and then the changing conditions of the overall trend of sedimentation will be considered together with the cyclic events.

From the frequently observed occurrence of marine beds closely following coal deposition the transgression of the sea over the low lying coal swamps must have been rapid, and was sometimes associated with reworking of the top of the underlying sediments, as at Astbury (section 3.1). The relative rise in sea level associated with the transgression was followed by quiet conditions in most areas and marine band shales with a goniatite-lamellibranch fauna were the normal deposit. In some near shore areas shelly faunas were developed,

as in the case of the Hd. proteus band at Astbury where there is evidence of current action depositing rippled sand under marine conditions.

In the Terrig River, Flintshire (Jones and Lloyd 1943 Loc. 38) below a horizon with E. aff. leitrimense and Ct. aff. nitidus (E2b2) a variable succession of shales with sandstones and some calcareous horizons has been seen by the author. Two coal bearing horizons are present, one of which has a ganister with rootlets below it. Some shelly marine beds are also present in this section and whilst the absolute age of these beds is uncertain they would appear to be within E2 and to represent nearshore conditions with occasional coal swamp development.

It was probably from such marginal marine areas with active sediment movement that the calcareous siltstones of the basin area were derived. No such sediments of E1-E2 age are preserved around the Staffordshire basin but their marginal situation would make them susceptible to erosion, and the fact that material was being moved into the basin from such an area indicates its unstable nature. Fig.5.3a shows the interpreted situation in north Staffordshire during deposition of marine band and calcareous siltstone sediments.

During deposition of the marine part of the cycle the land derived sediments begin to build out into the

Explanation Figs.5.3a-d

Diagrams to illustrate derivation of north Staffordshire basin cycles from an adjacent shelf area with delta development associated with periods of transgression.

M - Marine band.

C.S. - Calcareous siltstones.

Sh - Shale-mudstone.

P - Protoquartzites.

Fig.5.3a

Transgression and establishment of marine conditions on shelf and in basin. Deposition of marine band followed by calcareous siltstones derived from the marine, shallow water shelf deposits.

Fig.5.3b

Delta complex starts to build out over shelf and pro-delta clays cover shelf and fine material is carried into basin to form shale on top of calcareous siltstones.

Fig.5.3c

Delta complex reaches edge of shelf and sandy material from the delta front is introduced into the basin by turbidity currents originating at the front edge of the delta.

Fig.5.3d

Transgression begins and delta front retreats from shelf edge, or delta top is reworked, and sand supply to the basin stops resulting in deposition of a shale layer above the protoquartzite unit before a marine fauna is again established.

Fig 5.3a

TRANSGRESSION AND ESTABLISHMENT OF MARINE CONDITIONS ON SHELF.



Fig 5.3b

DELTA COMPLEX BUILDS OUT OVER SHELF.

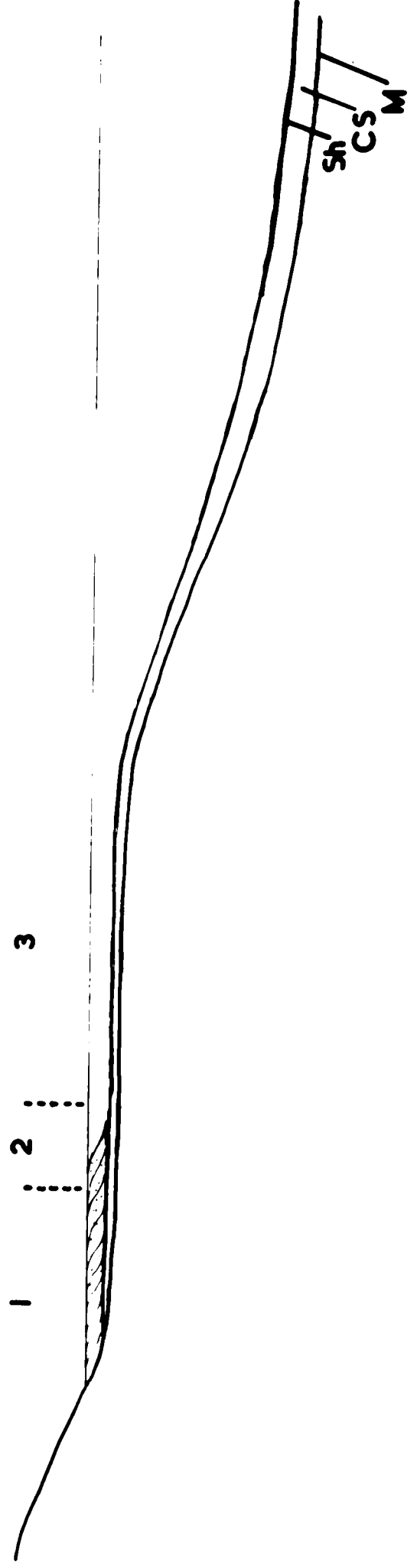


Fig 5.3c
DELTA REACHES EDGE OF SHELF.

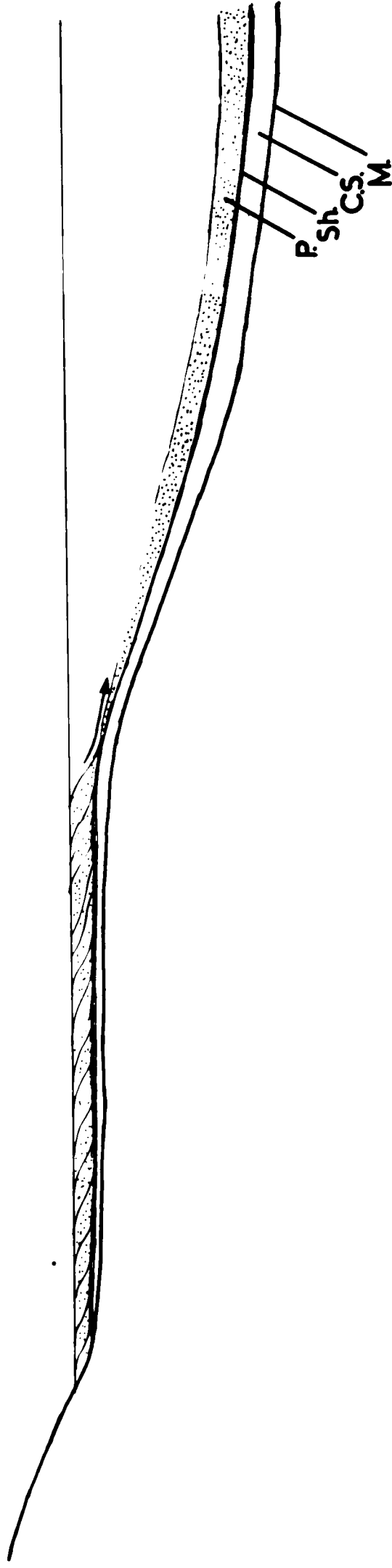
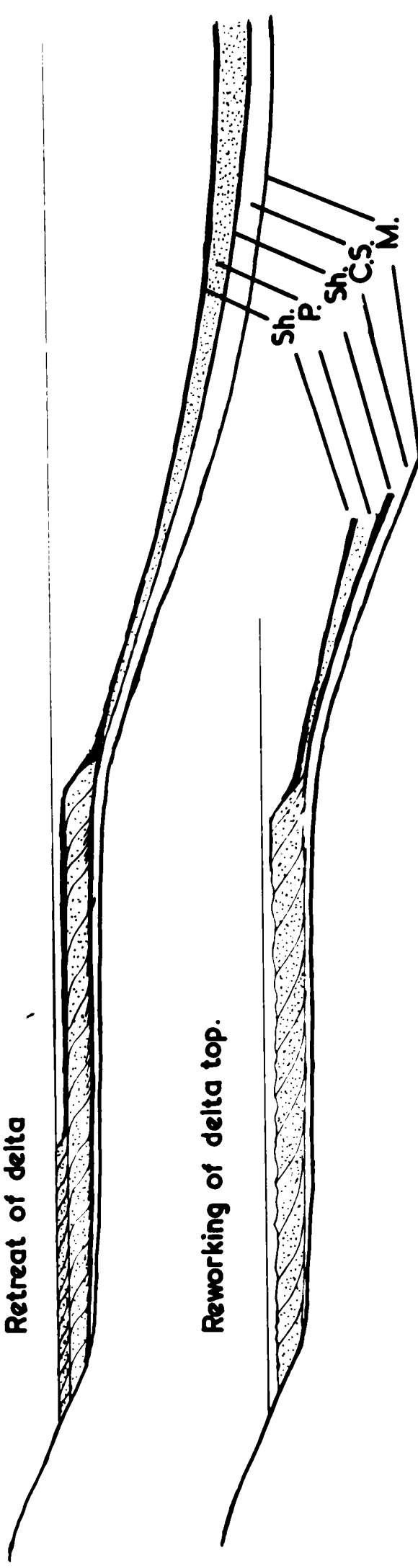


Fig 5.3d
TRANSGRESSION OF SEA OVER DELTA.

Retreat of delta



shallow coastal waters produced by the marine transgression. At any one point away from the coast the first material to be deposited will be the pro-delta clays and silts, followed by the sandy foreset beds of the delta which build the cycle up nearly to sea level, before the development of swamp conditions suitable for coal formation. Fig.5.3b illustrates the conditions during the gradual extension of deltaic deposits over the shallow water area produced by the marine transgression. During the seaward extension of the delta deposits, only the fine grained material carried in suspension from the coastal regions can reach the basin, and this fine material, grading landwards into the bottom set beds of the delta, covers the marine deposits on both the shallow coastal area and over the basin. It thus effectively halts calcareous siltstone deposition by covering both the supply and depositional areas with mud. This build out of the deltaic deposits is similar to that envisaged by Johnson (1961 p.326) in the case of the Yoredale cyclothems of northern England only the marine influence is less in the present area, limestone deposition being virtually absent.

The introduction of the land derived material to the basin area requires a slope on which to generate the turbidity currents which supplied the material to the basin. Generation of currents was probably by

slumping from an over steepened depositional slope possibly associated with submarine canyon formation. The presence of coarse material at the edge of such a slope requires the situation shown in Fig.5.3c where the foreset beds of the advancing deltas have reached the edge of the shallow water shelf area. Accumulation of material at the top of this slope produces instability with consequent downslope slumping and sliding leading to the generation of turbidity currents carrying material of deltaic origin into the basin.

The cessation of turbidite deposition in the basin area does not occur until a metre or so below the succeeding marine horizon. The cessation of turbidity currents could not be due to the complete filling of the basin, and hence elimination of the slope on which they were generated as there would then be nothing to prevent extension of deltaic conditions over the whole of the basin area - a feature which is clearly absent.

More probable is the theory that sediment supply in the source area of the turbidites was reduced until insufficient was being supplied for further generation of turbidity currents. This cessation of supply is interpreted as being the first indication of readvance of the sea over the low lying delta deposits. Transgression, or effective rise in sea level, would have had the effect of shifting the area of foreset sand deposition

rapidly shorewards so leaving only mud deposition from suspension as a record of the initial transgressive phase of the sea (Fig.5.3d). Eventually the reduction of the influence of land derived sediments would be such that marine conditions could once again be established and marine benthos flourish. In the shallow water area where the transgression was actively proceeding reworking of the underlying deltaic deposits undoubtedly took place by current and possibly wave action, but this did not occur in the quiet basin environment.

In the cycles recognised there is a progressive reduction in the fauna upwards within the cycle. The basinal marine bands contain thick and thin shelled goniatites, benthonic organisms, and abundant Posidonia individuals which were probably also benthonic (Holdsworth 1966 p.330). The calcareous siltstones contain rare thin shelled goniatites and seams of Posidonia species in the shales between the siltstone beds. The sideritic part of the succession lacks a shelly marine fauna but 'worms' were abundant as shown by their unroofed burrows on the bases of turbidite beds.

The apparent reduction in the 'marine' fauna upwards within the cycles could be due to a variety of causes. Preservation of calcareous fossils is obviously favoured in the lower, calcareous part of the succession, and the inferred slower rate of deposition would also lead to a

greater concentration of fossils than in the sideritic part of the succession. The absence of shelly fossils in the sideritic part of the succession could be due to a variety of reasons.

- 1) Non-preservation due to early diagenetic solution of CaCO_3 .
- 2) Deposition rate too high to allow shelly benthos to flourish.
- 3) Physical conditions of sediment/water interface unsuitable.
- 4) Chemical conditions of sediment/water interface unsuitable.
- 5) Chemical conditions of main body of water unsuitable.

It is clear that the first four of the above factors were different during protoquartzite deposition and calcareous siltstone deposition on the bases of diagenetic and physical features of the sediments already noted. It would therefore be surprising if a benthonic animal suited to the calcareous siltstone environment would survive the protoquartzite environment. The absence of the free swimming goniatites could be ascribed to the original low density of the goniatite population (cf. Holdsworth 1966 p.330), or to solution of any shells on reaching the accumulating sediment. It should be noted in this context that goniatites preserved crushed in marine band shale have often had their shells dissolved

out after crushing. If this solution preceeded crushing in a soft mud there would be little chances of final preservation in a shale.

Whilst it is tempting to relate the faunal change to reduction in salinity there is no unambiguous evidence as yet available on the actual salinities prevailing during deposition of the cycle.

Holdsworth (thesis 1963) considered that siderite deposition indicated conditions of reduced salinity due to mixing of run-off with the marine water of the basin. This reduction would have to affect a large area extending from Staffordshire to north of Edale, and thus the salinity of a very large volume of water would have to be altered. While this is possible there is no record of siderite forming at the present day in reduced salinities. The siderite is formed below the sediment/water interface, and hence the salinity of the water above this interface may have little bearing on the conditions required for siderite formation. The simplest solution is to relate both siderite formation and faunal change to alterations in sediment supply and diagenesis, but it must be admitted that environmental controls upon the specialised Upper Palaeozoic goniatite/lamellibranch faunas are still so little understood as to make any conclusion highly speculative.

The discussion so far has been restricted in that only the composite cycle of Coal Measures type has been compared with the events in the north Staffordshire basin area. It must be recognised that the composite cycle is not the same as the most frequently developed or modal cycle (Duff and Walton 1962 p.241) which may contain only a few of the lithologies making up the composite cycle. The modal cycle will reflect the general situation of the area. Thus in the Coal Measures, shale-seatearth-coal may be the most commonly occurring sequence, while in the Yoredale situation minor alterations of marine strata might prove to be modal. These variations from the ideal cycle must be taken into account as must also the 'minor cycles' recognised within major cycles by several authors.

Within the general pattern of cyclic sedimentation many authors have recognised minor cycles or rhythms, cf. Moore (1958, 59, 60), Johnson (1959, 61). Johnson (1959 Fig.3) shows several alternations of marine shale-shale-sandstone-rootlet bed-coal in the top part of his 'ideal' cyclothem. These 'minor rhythms' contain all the elements of the major rhythm except for a well developed and persistent limestone. There is obviously considerable variation in these minor rhythms as shown by Johnson (1959 Fig.4), and this extensive variation in rhythmic development is stressed by Duff and Walton

(1962) and by Duff et al. (1967 p.121). In the case of the sandstone members of the cycle, Moore (1960) shows the extensive local variations present due to channels and sheet sandstones derived from channels by crevassing of distributary banks. On the continent, various orders of cycles have been recognised with thicknesses of 4m., 10m. and 60m. (Van Leckwijck (1964) and discussion in Duff et al.(1967 p.143)), but as Duff et al. (1967 p.144) point out, the use of thickness to distinguish cycles is a bad practice since lateral variation in cycle thicknesses can frequently be demonstrated (Duff and Walton 1962). Wanless (1964 p.604) stresses the presence within Pennsylvanian cyclic sediments found in the eastern and central United States of both local and widespread cyclic units and considers that the local events are due to the introduction of clastic wedges which may occur several times in a widespread cycle at a particular locality.

There do appear to be grounds for believing that superimposed on major cycles of sedimentation minor oscillations occur, less persistent in time and space than major cycles and involving only partial sequences of the major cycle. The extreme irregularity of major cycles and the frequently admitted non-development of parts of the cycle point to minor cyclic influences being superimposed on a major trend which is itself

part of a larger cycle.

It has been previously argued that minor cycles occur within the basin deposits both of calcareous siltstones and protoquartzitic turbidites, the thicker beds of both these lithofacies being grouped together and separated by predominant shale. This tendency has been noted in turbidite formations on greatly varying scales. Megarhythms in the Carpathian Flysch hundreds of metres thick with coarse beds at the base dying out upwards have been recognised by Ksiazkiewicz (1960). In the Aberystwyth Grits the thicker greywacke turbidite beds are grouped together and separated by thin bedded greywackes, the variation taking place over a few metres (Wood and Smith 1959). In this case a group tends to start with a thick bed and the beds then reduce in thickness upwards. In the Ordovician of the Rhinns of Galloway (Kelling 1961) cycles of about 10m. terminate upwards in a thick bed. Kimura (1966) has constructed sandstone/shale diagrams of turbidite formations from the Tamagawa and Oigawa districts of Japan and found major cycles with superimposed minor cycles of the order of 10m. He recognised types of cycle with upward increase in turbidite bed thickness.

While it seems established that in turbidites there may be a minor cyclicity present it is difficult to relate such cycles to any particular cause. Possibly

the cycles are developed due to irregular downwarp of a basin area or irregular movements on faults in a basin of deposition as envisaged by Neef (1964) in the Pliocene of New Zealand. Any explanation of the minor cycles must account for the supply of sediment to the area in which the turbidity currents were generated. It is quite possible that the minor cycles represent fluctuation in supply of material rather than tectonic movement of the basin. The two secondary modal rhythms of Duff and Walton (1962) each containing sand and coal conditions as seen in a shallow water Coal Measures environment could control such sediment supply to the edge of a shelf area giving rapid sand supply for a short period favouring the generation of turbidity currents on a slope due to oversteepening of slope deposits. The change to marsh conditions bordering such a slope would reduce the sandy sediment supply. Stability of the slope might be established and only fine grained material deposited in the basin area.

There is no evidence that this has in fact happened in the present area but it is worth considering that minor changes in sedimentation patterns marginal to the basin could have a great effect on sediments in the basin area, particularly when it seems probable that the basin deposits of north Staffordshire are derived from a shallow water deltaic area.

The foregoing discussion has been restricted to a comparison of the basin cycles observed with those known to occur in shallow water, nearshore environments at the same time. No attempt has been made to explain the overall trend in sedimentation from calcareous to sideritic sediments recognised in Section 5.1 in combination with the cyclic events.

The foregoing description of the possible mode of deposition of the cycles in the basin requires that periodic changes in relative sea level must take place to produce the transgressions represented by the marine bands. There is ample evidence that this model of events is far too simple to account for all the features of the Carboniferous basin areas. It has already been shown that great differences in thickness of sediment of the lower Namurian and upper Viséan exist in the region especially between the P and E sediments of the massif area, e.g. Ashover, and the 'gulf' or 'basin' areas described in Part 4. It is in fact at this level in the succession that the greatest difference exists between massif and basin sediment thicknesses. The ultimate extension of this difference is the presence of an unconformity in this part of the succession cutting out parts of E or P. Such an unconformity has already been described from around the margins of the north Staffordshire basin area, and from other areas (see Part 4).

Clearly, mutual relative movements of basin, massif and sea level were taking place at this time. Basin areas were subsiding and receiving their thick sediments, while the massif areas subsided more slowly relative to sea level and received thinner sediments, or may have risen with resulting condensation of the succession or actual unconformity when emergence of the massif areas occurred.

In the Staffordshire-Derbyshire area relative movement appears to have taken place by flexuring at the basin margins possibly associated with faults at depth. The west side of the north Staffordshire basin may also have been controlled by faulting which produced the great differences in thickness between the P and E sediments of Astbury and Lask Edge. Further north the Craven faults were active in separating the Askrigg Block from the basin area to the south.

These tectonic movements had different effects in the different areas of basin and massif described in Part 4, and while they are possibly the major feature responsible for local sedimentation changes they do not account for the widespread marine bands which are most likely to be due to sea level changes.

There is, however, through the British Carboniferous, a general change from Carboniferous limestone deposition to coal measures deposition and the result of intra-

Carboniferous movements was one of emergence which culminated in the Permian. The part of the succession studied here marks the important point in the local succession when general calcareous deposition ceased in response to the increasing amount of land derived material being deposited in the basins. The way in which the calcareous siltstones and protoquartzites fit into this general plan can be seen by considering the sequence of lithologies present in the local Carboniferous.

After completion of the transgression of the Lower Carboniferous, carbonate deposition was established over a wide area of the British Isles in response to the warm climate and shallow shelf sea. The transgression resulted in a relative lowering and reduction in area of the remaining land areas - such as Wales - and hence detrital supply to the seas was greatly reduced, and consequently in the Lower Carboniferous the coastal strip of land derived clastics is usually very narrow. During the Lower Carboniferous the basin and massif areas of deposition became established and Carboniferous 'reefs' flourished at the massif margins and marginal to existing land areas (e.g. Flintshire) from which detrital supply must have been practically nil. With the initiation of strong relative movements of basin and massif areas in mid- and late Viséan and early Namurian times more detritus was eroded from the land areas and carried into the seas forming the mixture of land derived

clastics and carbonates found in north Wales and the Formby Borehole and from which the calcareous siltstones of the Staffordshire and north Derbyshire basins are thought to be derived.

With further raising of the land areas local climatic conditions probably changed with increase in rainfall and run-off and the development of extensive vegetation. Conditions then became established for the formation of deltas and low lying swamp areas bordering the coastline which formed the supply source of the protoquartzites. Supply of land derived clastics was now so rapid as to favour the production of diagenetic siderite, because of a low concentration of sulphur species within the sediment.

Even if sediment surface conditions were suitable for the precipitation of calcite, the high proportion of dissolved iron in the deltaic material, combined with a low concentration of dissolved sulphur species, would favour its alteration to siderite during diagenesis as was found by Taylor and Spears (1967) in the case of the carbonate bed associated with the Mansfield Marine Band. There seems to be a general rule that detrital sediments derived from deltaic areas, and probably deposited relatively quickly, are characterised by diagenetic siderite due to low concentrations of dissolved sulphur species produced by sulphate reducing bacteria. A high concentration of such bacteria and sulphur species is

produced diagenetically during slower deposition. Under such conditions the available iron is incorporated into pyrite while calcite and dolomite carbonates are produced. In the calcareous siltstones these minerals are found, but some iron is also incorporated in ankerite. There is no real evidence that the variations are due to salinity changes since siderite can be found associated with marine faunas as in the G. subcrenatum band and in brachiopod bearing beds above the Hd. proteus band.

The cycles already described show alternations in sideritic and calcitic sediments caused by variations in type and rate of detrital supply. The changes in sedimentation are probably due to changes in sea level superimposed on the general trend to increase the proportion of land derived material. By E2a2 time the mixed calcareous-clastic supply area of the calcareous siltstones had probably been permanently covered by land derived clastics of deltaic type and as further extension of land areas continued until the Permian, conditions for the production of the calcareous siltstones were never re-established.

It is of interest to note that in north Wales the Carboniferous limestone is succeeded by the very variable Cefn-y-Fedw sandstone, which contains a mixture of shelly carbonate and clastic deposits, and that near the top of this sandstone a coal and ganister can be seen in the

Terrig River below the E2b1 horizon with Ct. aff. nitidus and E. aff. leitrimense - thus lending a little support to the above theories.

5.4

POSSIBLE CAUSES OF THE CYCLIC SEDIMENTATION OBSERVED.

Duff et al. (1967 Ch. 11) have reviewed the extensive literature on the various theories that have been advanced for the origin of cyclic sedimentation. They review briefly the sedimentary, tectonic, eustatic and climatic theories which have been advanced to explain cyclothems differing very widely in type, extent and environment. In the Carboniferous it is necessary to treat each case of observed cyclicity of sediments on its own merits and not to attempt a general solution to the whole problem, since there may be no general solution. The great variability of Carboniferous cycles points to the possibility that several mechanisms of control may be in operation.

Duff and Walton (1962) considered that Coal Measures cycles in the east Pennines were dominantly due to environmental changes of a local nature and of sedimentational control. The cycles varied to such an extent over the area they studied that correlation was sometimes difficult. They do however recognise three cycles containing the Pot Clay, Clay Cross and Mansfield Marine Bands which are 'significantly different from all other cycles in thickness and extent', and can be traced throughout Britain and Europe. These cycles they consider might be the result of regional diastrophism, or world wide

changes in sea level.

The marine horizons in the succession studied are mainly of this type in being of wide areal extent, being recognised clearly in Eire, England and on the Continent of Europe (see palaeontology section). Marine horizons that may be only of local extent are those occurring in the shales between the top of the Unit C protoquartzites and the Ct. edalensis horizon and also the described horizons above Ct. edalensis. The actual Ct. edalensis band is of wide extent.

The mechanism controlling the deposition of these bands must be of great areal effect, influencing the British Isles and Europe. Local sedimentational control is not the solution in this case — as is shown by the widely differing successions of Slieve Anierin (Yates 1962), Edale, north Staffordshire, the north of England (Wilson and Thompson 1959, Wilson 1960a) and Ashover (Ramsbottom et al. 1962). The development of the distinctive marine bands in widely different basin and massif areas of sedimentation means that the control mechanism must have been synchronous over a wide area and of sufficient magnitude to affect areas of very different setting. During E1-E2 times tectonic movements certainly took place resulting in unconformities in north England (Wilson and Thompson 1959), the Hope area in Flintshire (Jones and Lloyd 1943 p.254) and around

the margins of the Derbyshire massif (e.g. Upper Dove and near Edale, Derbyshire). While unconformity is found at these places other localities show a reduced sequence with little or no sedimentation, as at Ashover, Derbyshire, (on Derbyshire massif) where thin phosphatic shales occur in E1 and the base of E2 (Ramsbottom et al. 1962) and marine bands are not developed. At Astbury, in the area described in this thesis, a relatively thin succession represents E1 and E2 and the characteristic marine horizons are absent (Section 3.1). Conversely, within the north Staffordshire, and other basin areas thick deposits represent E1 and E2, and considerable relative subsidence of the basin must have taken place at this time. It can therefore be demonstrated that differing rates of uplift and also of downwarp were operating during E1 and E2 times in England in different places, and thus it seems impossible that the development of extensive synchronous marine bands is due to widespread tectonic events as envisaged by Bott and Johnson (1967). Such events would have to be extremely widespread, and effective over both the massif and basin areas where the marine bands are developed, and faulting in the upper crust can hardly be expected to have such even effects over such large areas as pointed out by Benfield (in discussion of Bott and Johnson 1967 p.439) — especially when the foregoing evidence of local contemporaneous differential movement is considered.

Westoll (1962) combined the effects of compaction, isostasy and tectonic forces in an attempt to account for the Yoredale cyclothems. While appearing feasible for the case he chooses, it cannot account for the widespread marine horizons of E1-E2 which are separated by widely different lithologies and thicknesses in widely varying settings relative to basin and massif areas.

It is considered that the development of the widespread marine bands must be due to an equally widespread and synchronous change in conditions resulting in transgressions of the sea. The simplest event needed to cause the required transgression is a marked rise in sea level. Whether this is caused by glacial control (cf. Wanless and Sheppard 1936) is open to speculation but the presence of Carboniferous continental glacial episodes in South America, South Africa, India and Australia (King 1958) lends considerable support to the glacial control theory.

Wanless (1960) recognised numerous alternations of glacial and non-glacial sediments in Victoria, Australia, and concluded that there were many minor glacial episodes in Australia within the Upper Carboniferous-Permian glaciation. In one section, 51 alternations of glacial and non-glacial sediments were noted. The problem of course does not end here since a mechanism

capable of producing the various glaciations required must now be found. Some form of climatic control is possible, but the evidence at present available regarding world wide climate variations with time, their cause and effect on the sedimentary environment, is insufficient to warrant further consideration here (see Duff et al. 1967 p.248-250).

It is concluded that the cycles observed, with their overall trend from calcareous siltstones to protoquartzitic turbidites derived from a deltaic area, are due to intermittent rises in relative sea level superimposed on a general trend of uplift of land areas relative to the sea. Many strong local movements, principally connected with the margins of massif and basin areas, produced successions with great local variation as seen in and around the north Staffordshire basin area.

PART 6.Locality List with Grid References.

<u>Locality</u> <u>No.</u>	<u>Grid Ref.</u>	<u>Horizon</u>	<u>Notes</u>
HURDLOW			
001	02896042	E2a1	
002	02716073	E2b1	<u>L. longirostris</u> - <u>Ct. edalensis</u> - section in bank on S. side of small tributary gully
002A	02716074	E2b1	Bentonite B7 upper <u>C. cf. C.</u> <u>subplicatum</u> horizon - section in bank on N. side of small tributary gully
003	02736085	E2b1	Bentonites B5, B6 to <u>Ct. edalensis</u>
008	02876069	E2a2	top of bank
008A	02916069	E2a1	base of bank
009	02876082	E2a2	at stream level
010	02776085	Unit C	Protoquartzites
011	03066056	E1c + Unit A	
012	03056060	E1c + Calc. silts.	Unit β
013	03016082	Base of Calc. silts.	β
014	03146081	Top of Calc. silts.	α
015	02756068	Unit C	
016	03716124	E1	Unit A - Calc. silts. β - Unit B
017	03806078	Unit A	
018	02716077	E2b1	bullions above <u>Ct. edalensis</u>
019	03716115	Unit A	

020	02876082	E2a1
021	02916087	E2a1
022	02966106	E2a2

UPPER SHIRKLEY

054	92105907	E2a2 + Unit C downstream
055	92065906	E2a1

DALEHOUSE WOOD (BLACKWOOD HILL)

060	91515602	E2b1 <u>Ct. edalensis</u>
061	91905548	Unit C
062	92285544	Unit C
063	92435546	Unit C ?
064	92465545	Unit C ?
065	92615545	Unit C ?
066	92815540	E2b3 Cherts

SPRINK

070	91685780	Calc. Silts. Unit β
071	91655786	Unit A
072	91865783	Unit B
073	91915783	Unit B
074	92095784	E2a2
075	92705779	Shale-Mudstone - unknown horizon

BROADMEADOWS

080	92075858	Units B or C
081	91835857	Calc. Silts. and shale with <u>P. corrugata</u> near E1c

PYECLOUGH

101.	04656488	E2a2 <u>E. sp.nov.II, L. longirostris</u>
102	04386478	E1c - Unit A
103	04436468	Unit A
104	04476467	E1c
105	04486470	Calc. silts. Unit β
106	04556472	E2a1 - Unit γ
107	04676494	E2b1 <u>Ct. edalensis</u>
108	04586473	E2a2
109	03686483	Calc. Silts. Unit α ?
110	03566483	Unit B
113	04696495	E2b2 <u>Ct. cf. Ct. nitidus</u>
114	04676496	Bentonite B7
115	04686497	E2b1 ? Bullion in shale
116	04696500	E2b3 cherts
120	03506484	E2a1

OAKENCLOUGH

151	04766323	Unit A
152	04806323	E1c
153	04816323	Unit β
154	04846324	Unit B
157	04916340	E2a2 (poor exposure)
158	04946341	Unit C
159	04996345	E2b1 <u>Ct. edalensis</u>
160	05056350	E2b3 Cherts
161	05016368	E2b3 Cherts
165	04806300	E1c

BLAKE BROOK

201	05726088	<u>L. longirostris</u> - <u>Ct. edalensis</u> E2b1
202	05786094	E2b1 <u>C. cf. C. subplicatum</u> - <u>Ct. edalensis</u> - Bentonite B7 (cliff section)
203	05886098	E2b1 <u>C. cf. C. subplicatum</u> - <u>Ct. edalensis</u>
204	05606077	Unit B
205	05646080	Unit Y
206	05656082	Shale-Mudstone + Siderite
207	05626078	Unit B
208	05606075	E1c
209	05646078	E2a1
210	05666085	E2a2
211	05666097	E2a2
212	05736078	E2b1 <u>Ct. edalensis</u>
213	05936102	E2b1 <u>Ct. edalensis</u>
214	05906100	E2b1 <u>Ct. edalensis</u>
215	05576104	E1c

SUNNYDALE

251	04166603	E2a2 - Unit C
252	04006605	E2a1
253	04096601	E2a2

HAYES FARM, BUTTERTON

260	06425728	E1a <u>C. cf. C. leion</u>
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UPPER DOVE

301	07536685	E2a2
302	07276690	E2a2
303	07056695	E2b1 <u>Ct. edalensis</u>
304	07736676	E2b3 Cherts
305	07446688	E2a2
306	07236690	Unit C equivalent

CROKER HILL

351	93416683	E2a2
352	93386675	E2a2
353	93306669	E2a1
354	92806723	E2b1 <u>Ct. edalensis</u> - Bentonite B6
355	93606682	Unit C
356	94056761	E2a2
357	93836844	Unit 8
357A	93796842	E2a1
358	93436858	E2a1
359	93566868	Unit B
360	93606945	Unit B ? Fox Bank Quarry
361	92596741	H - R
362	94136781	E2a2
370	92206938	Slump beds, Hawkshead Quarries
371	92226942	Massive sandstones, Hawkshead Quarries
375	92796882	Slumped ssts - horizon unknown

HIGHER MINNEND

381	93696455	Unit C
382	93746459	<u>L. longirostris</u>
383	93796462	E2b1 <u>Ct. edalensis</u>

GUN HILL

401	96456030	E2a2
402	96506175	Unit C Gun Stone Pits

OLD HAG

410	97596315	E2b1 <u>Ct. edalensis</u>
411	97566310	Bentonites B4, B5, B6

THORNCLIFF

421	01765859	Bentonites B5, B6
422	01825858	Unit C
423	01975861	Unit B
424	02135863	Unit A
425	02225862	Unit A

WIGGENSTALL

441	08946091	E2b3 cherts - in bridge foundations
444	09066079	E2b3 cherts
445	09096078	Ssts. + siderite above cherts
446	09086060	E2b1 <u>Ct. edalensis</u>
447	09056055	<u>L. longirostris</u> - top of Unit C

EASING

461	01925785	E2b1 <u>Ct. edalensis</u> . <u>L. longirostris</u> 30 yds upstream in bank
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462	01975785	Unit C
463	02115786	E2a2
464	02155788	Unit B
465	02255789	Unit β

WARSLOW and ELKSTONES

501	06435901	E2a2
502	06455888	E2a2
503	06455881	Unit γ
504	06465878	E2a1
505	06435870	Unit B - E2a1
506	06425862	Unit β
507	06385856	E1c Unit A - Unit β
508	06385851	junction Unit A - Unit α
521	07165845	Unit A - E1c - Unit β
522	07515887	Unit C
523	07505880	E2a2
524	07525872	Unit γ - Sh.-Mdst. + Siderite - E2a2
525	07535867	E2a1 (poor)
526	07505824	Calc. Silts. (unit unknown)

ASTBURY LIME WORKS

540	86105890	Calc. Siltstones
541	86235923	Astbury sandstone
542	86125888	Calc. Siltstones + <u>Lingula</u>
543	86175887	Fauna E2b ?
544	86255930	Tuff + Corals, inc. <u>P. murchisoni</u>
545	86955930	<u>Hd. proteus</u> in ganister quarry

WATERHOUSES

561	07355024	E2a2
562	07225038	E2b Bullions with <u>P. corrugata</u>
563	07205041	E2b Bullions with <u>A-D</u> sp. in shale
564	06825042	Shale-Mudstone with siderite
565	07585016	Shale-Mudstone, <u>P. corrugata</u> ; <u>A-D</u> sp. (Horizon unknown)

WINKHILL - MARTINSLOW

581	06625204	E2a1 - Unit γ , Sh.-Mdst. + Siderite
582	06705265	Unit α
583	06775262	E1c
584	06695250	Protoquartzites - Calc. Silts. transition, unit unknown
585	06685229	Sh.-Mudst. + <u>P. corrugata</u>
586	06725218	E1c
587	06685215	Unit

WATERHOUSES SCHOOL SECTION

601	08225017	<u>E. pseudobilingue</u> E1b
602	08265015	Bullion horizon - unfossiliferous, Low E1 ?

WATERHOUSES RAILWAY

621	07334968	E2a1 ? (Morris Loc. 10)
622	06804974	E2a2 (Morris Loc. 33)
623	077-497-	E1a shales and bullion horizons (see Morris 1967)

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AREAL DISTRIBUTION OF CALCAREOUS SILTSTONES

Fig 2.4Ga

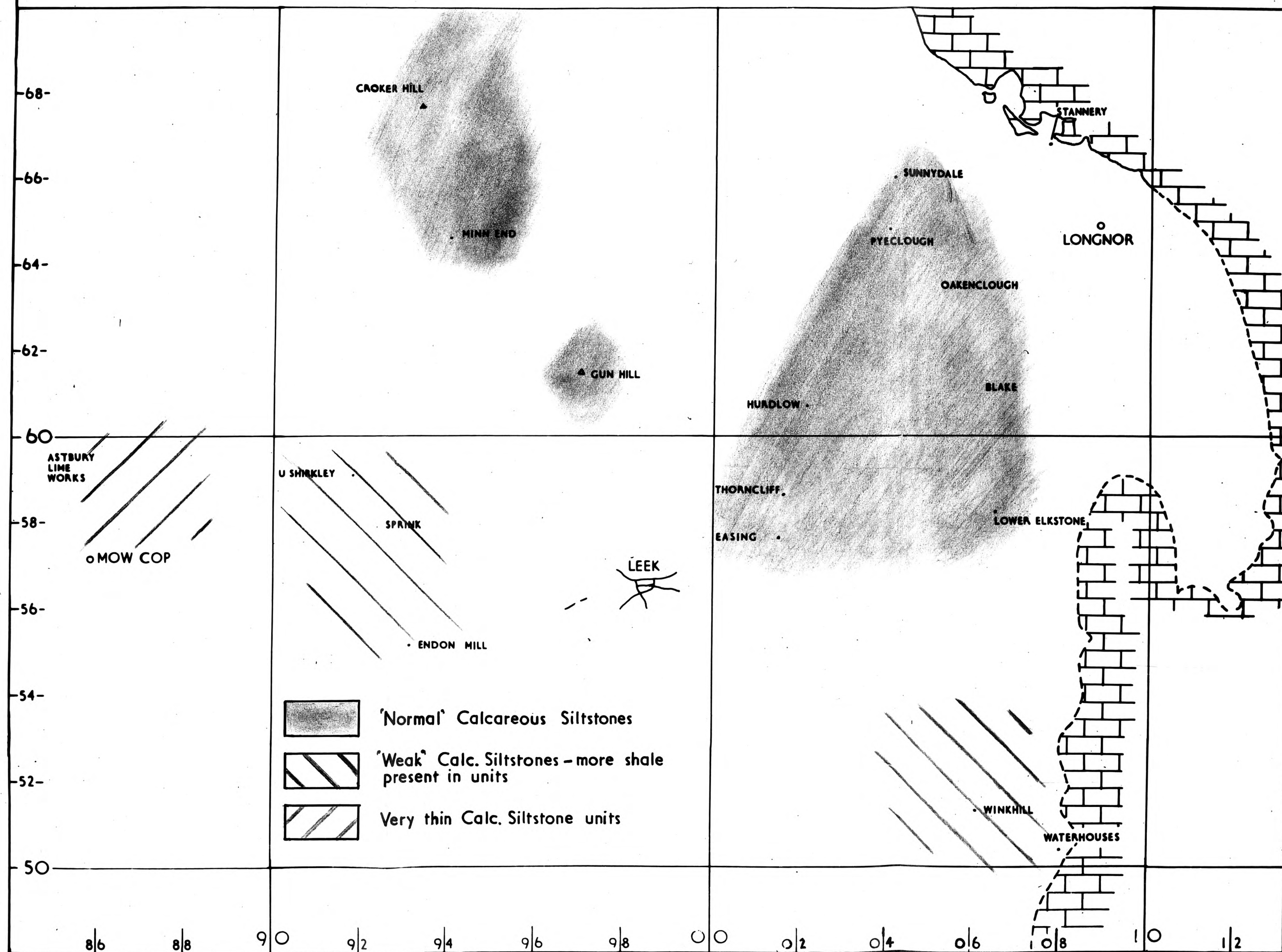
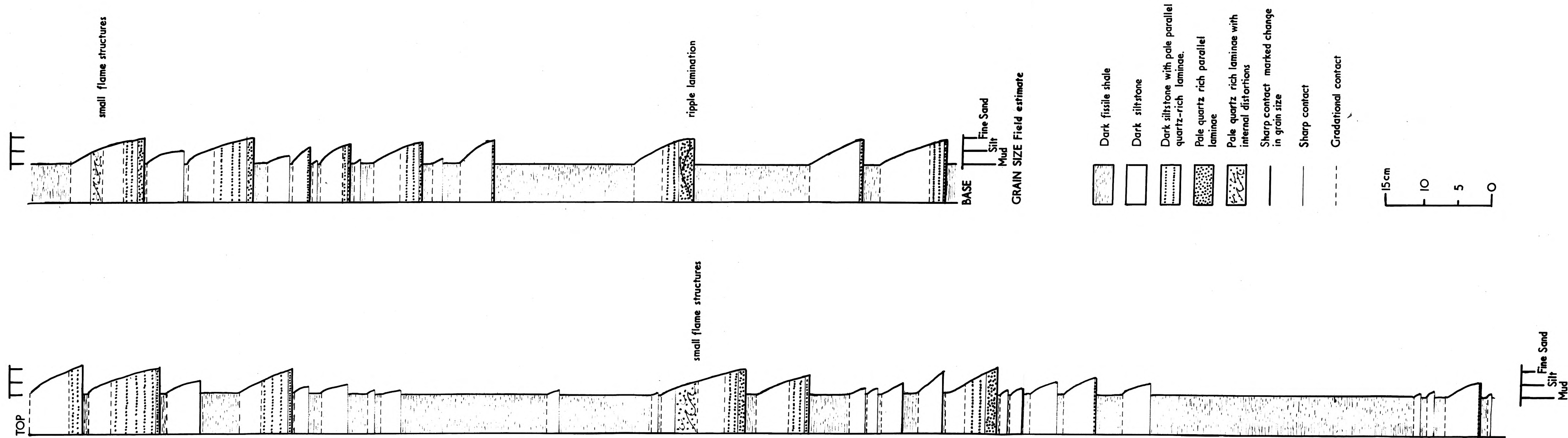


Fig 2.4Gb

LOG. CALCAREOUS SILTSTONES.
Unit β Loc.153.



2 • FORMBY No 1
380 m ? E
CONTAINS CHERTS, DOLOMITES, SHALES,
LIMESTONES, CALC. SANDSTONES, SSTs,
SOME SHELLY MARINE FAUNAS

Manchester

Holywell •
E. bisulcatum
IN HOLYWELL Sh.

G RESTS ON
Lr CARB

Mold

E. bisulcatum
IN Cefn-y-Fredw
Sst. WITH SOME
SHELLY MARINE
SANDS

WESTERN MARGIN
OF BASIN

ASTBURY
P AND E THIN
COALS IN P&H

WESTERLY LIMIT OF
E ZONE TURBIDITES

G RESTS ON
CARB LST

U COAL MEASURES
ON Lr CARB.

NAMURIAN THIN
OR ABSENT

U COAL MEASURES
ON PRE-CAMBRIAN

KEELE BEDS ON
LONGMYNDIAN

Macclesfield

E AND P THICKEN NORTH
FROM UNCONFORMITY

Buxton
E UNCONFORMABLE
ON VISEAN REEFS

N. STAFFS. BASIN

E THICK
TURBIDITES
FROM SOUTH

THICK
PENDLESIDE
+ VORDALES

Stoke

Market Drayton

Newport

COAL MEASURES
ON
PRE-CAMBRIAN

Wolverhampton

COAL MEASURES
ON DEV. AND SIL.

EDALE GULF

2500' - 760 m

2000' - 608 m

Sheffield

1500' - 456 m

GAINSBOROUGH TROUGH

Gainsborough

THIN E-R, SHALES
THICKEN TO 200m
IN WEST WITH
MUDDY LIMESTONES

Lincoln

MILLSTONE GRIT
UNCONFORMABLE
ON ? S₂

R, RESTS ON
UPPER VISEAN

VERY THIN
SHALES ON
MASSIF

Lr NAMURIAN
THIN

MASSIF MARGIN

EASTERLY LIMIT OF
E ZONE TURBIDITES

THICK E WITH
GRADED SSTs

WINDMERPOOL GULF

Nottingham

RAPID SW EXPANSION
OF E AND P INTO GULF

Grantham

E AND P THIN
OR ABSENT

VERY THICK
E. AND P

NAMURIAN RESTS
ON PRE-CAMBRIAN

SOUTHERN MARGIN OF GULF

Leicester

MODERATE TH.
E AND P SHALE

THIN MILLSTONE GRIT
AND Lr CARB. NO E.

AREAL SITUATION OF N. STAFFS BASIN

Fig. 4a

Areal situation of north Staffordshire area during E1-E2 times.

Names of numbered boreholes and sources of information.

1.	Holywell (borehole and area)	Jones and Lloyd 1943; Wood 1936
2.	Formby No. 1	Kent 1948
3.	Ternhill	Wills 1956
4.	Stoke-upon-Tern	"
5.	Egmond	"
6.	Bowsey Wood	Earp and Calver 1961
7.	Apedale	Giffard 1923
8.	Werrington	"
9.	Gun Hill	Hudson <u>in</u> Hudson and Cotton 1945a
10.	Edale	Hudson and Cotton 1945
11.	Alport	Hudson and Cotton 1943
12.	Chartley	Stevenson and Mitchell 1955
13.	Whittington Heath	Mitchell 1954
14.	Ashby	Mitchell 1941
15.	Wardlow Mires	Inst. Geol. Sciences Ann. Report 1966
16.	Uppertown	Ramsbottom, Rhys and Smith 1962
17.	Highoredish	"
18.	Tansley	"
19.	Calow No. 1	Smith et al. 1967
20.	Ironville No. 2	"

21. Duffield	Inst. Geol. Sciences Ann. Report 1966
22. Hathern	Falcon and Kent 1960
23. Windmerpool	"
24. Long Clawson	"
25. Colston Bassett	"
26. Langar	"
27. Plungar	"
28. Sproxton	"
29. Bottesford	"
30. Claypole	"
31. Thorpe	"
32. Eakring	"
33. Sutton-on-Trent	Edwards 1967
34. Tuxford	"
35. Bothamsall	"
36. Nocton	Falcon and Kent 1960